Miscellaneous Paper CERC-95-3
February 1995

US Army Corps
of Engineers
Waterways Experiment
Station

Literature Review on the Geologic Aspects of Inner Shelf Cross-Shore Sediment Transport

by J. Bailey Smith

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Prepared for Headquarters, U.S. Army Corps of Engineers

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Final report

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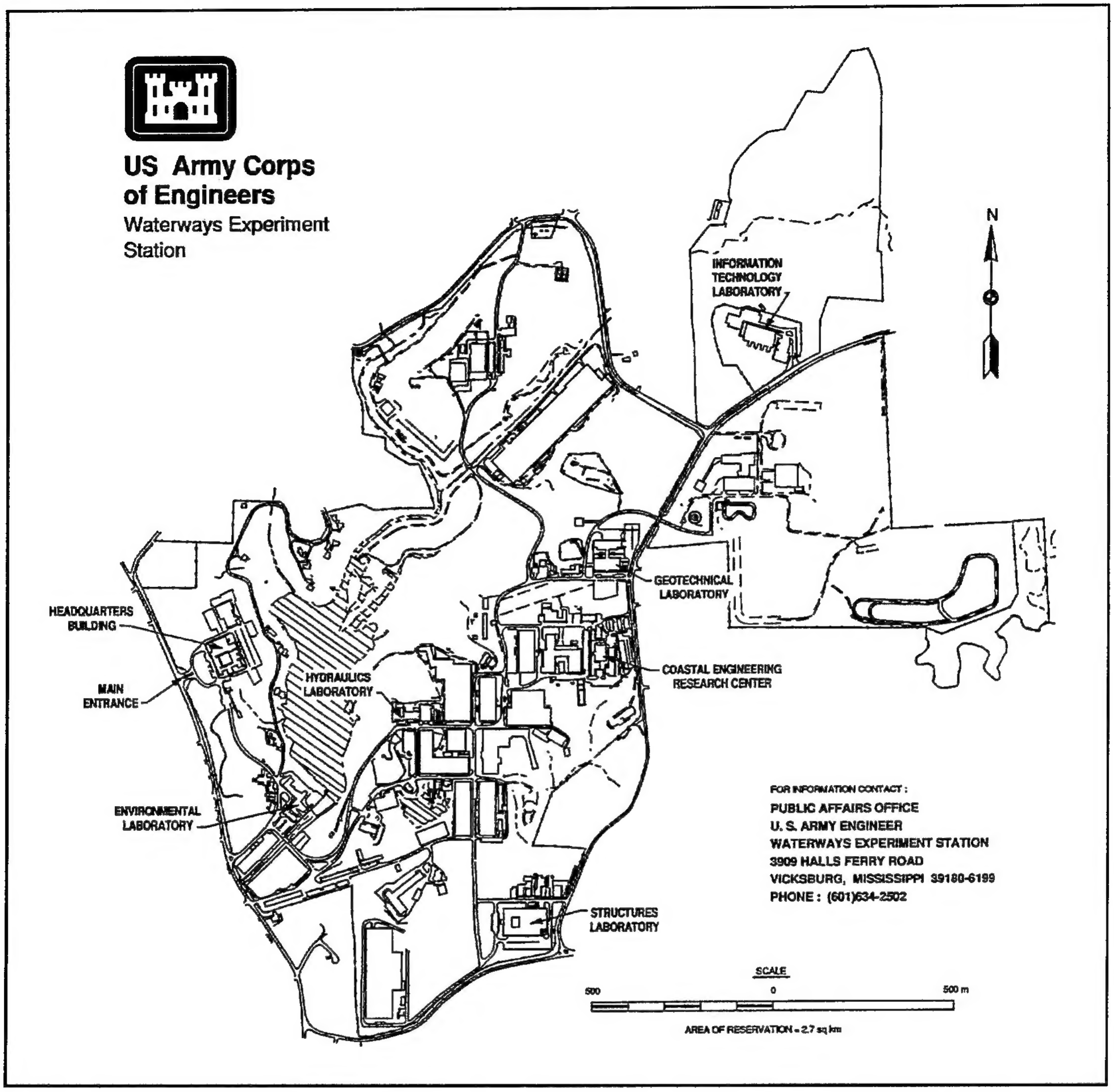
Prepared for U.S. Army Corps of Engineers
Washington, DC 20314-1000



Waterways Experiment Station Cataloging-in-Publication Data

Smith, J. Bailey.
Literature review on the geologic aspects of inner shelf cross-shore sediment transport / by J. Bailey Smith ; prepared for U.S. Army Corps of Engineers.
161 p. : ill. ; 28 cm. — (Miscellaneous paper ; CERC-95-3)
Includes bibliographic references.
1. Continental shelf — Sediment transport — Bibliography.
2. Continental shelf — Physical geography. 3. Geology, Stratigraphic.
4. Sediment transport — Bibliography. I. United States. Army. Corps of Engineer. II. U.S. Army Engineer Waterways Experiment Station.
III. Coastal Engineering Research Center (U.S.) IV. Title. V. Series: Miscellaneous paper (U.S. Army Engineer Waterways Experiment Station) ; CERC-95-3.
TA7 W34m no.CERC-95-3

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Preface

The study reported herein results from research performed by the U.S. Army Engineer Waterways Experiment Station (WES), Coastal Engineering Research Center (CERC) under the Geologic Analysis of Shelf/Beach Sediment Interchange Work Unit 32821, Coastal Geology and Geotechnology Program, authorized by Headquarters, U.S. Army Corps of Engineers (HQUSACE). Mssrs. John F. C. Sanda, John G. Housley, and John H. Lockhart were HQUSACE Technical Monitors. Ms. Carolyn Holmes was CERC Program Manager.

This report was prepared by Mr. J. Bailey Smith, Coastal Geology Unit, Coastal Structures and Evaluation Branch (CSEB), Engineering Development Division (EDD), CERC, under the general supervision of Mr. Thomas W. Richardson, Chief, EDD, and Ms. Joan Pope, Chief, CSEB. Director of CERC was Dr. James R. Houston, and Assistant Director was Mr. Charles C. Calhoun, Jr. Mr. William A. Birkemeier and Mr. Andrew Morang, EDD, provided suggestions as to the content of the report.

At the time of publication of this report, Director of WES was Dr. Robert W. Whalin. Commander was COL Bruce K. Howard, EN.

1 Introduction

This literature review addresses sediment transport across the inner portion of the continental shelf, also referred to as the shoreface, or, as in this report, the *inner shelf* (Figure 1). The inner shelf extends from the seaward edge of the surf zone to the landward edge of the continental shelf. It is affected by the strong agitation that results from sediment resuspension caused by shoaling of nonbreaking waves. The inner shelf is friction-dominated by both bottom and sea-surface boundary layers which overlap and frequently occupy the entire water column (Wright, in press). The inner shelf differs from the surf zone, which is also characterized by strong agitation of the bed by waves. The bed of the surf zone, however, is affected by the bore-like translation of waves following wave breaking (Komar 1976), and by wave-induced longshore currents and rip currents.

Cross-shore transport of sediment across the inner shelf has a great effect upon short- and long-term fluctuations of beach and surf zone sand storage as well as the morphology and stratigraphy of the inner shelf. Although surf zone and nearshore processes and sediment transport have been extensively addressed in the literature, inner shelf processes and sediment transport, particularly in the cross-shore direction, are not well understood. The complexity and interdependence of the mechanisms controlling transport on the inner shelf make it very difficult to comprehensively understand and describe the processes affecting sediment on the inner shelf. In response to this, Wright (1987) stated that a goal of the scientific community should be "to devise a more universal conceptual framework capable of better accounting for shoreface transport, erosion, and deposition in time and space."

Knowledge of sediment transport to and from the inner shelf region has important implications to engineering works such as beachfill design and dredged material placement. In computing a sediment budget for a beachfill project, offshore gains and losses are usually assumed to be negligible in the sediment budget calculations. While this assumption recognizes the difficulty in quantifying inner shelf exchanges, it is probably incorrect during significant events. Defining limits for the active nearshore profile under varying conditions can aid in placing dredged material so that it will likely move onshore, offshore, or remain stable.

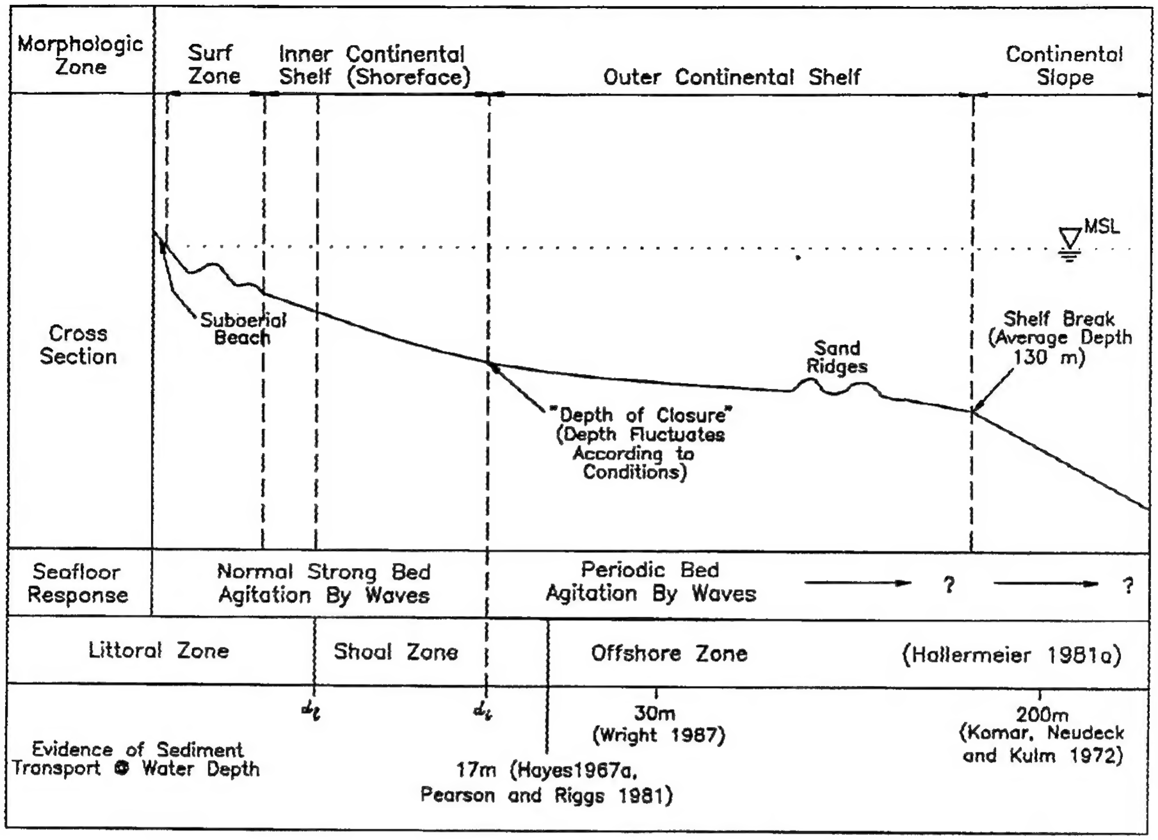

Figure 1. Continental shelf cross-sectional profile (site specific to the mid-Atlantic Bight of the United States). D_l and d_i (from Hallermeier (1981a)) refer to the seaward limit of surf-related effects, and the seaward limit to sand motion by normal waves, respectively

Most models predicting shoreline change and cross-shore profile shape and changes are based on a *profile of equilibrium* which recognizes that for a given wave condition or average wave condition there is a profile shape (concave upward) that is in equilibrium with the wave conditions. While useful, this concept ignores the fact that, in addition to wave action, many other processes affect sediment transport. Moreover, cross-shore sediment transport is also affected by the regional geological framework and profile shape, as well as hydrodynamic conditions.

Purpose

The purpose of this report is to summarize literature which addresses the exchange of sediment between the beach and the inner shelf through analysis of physical processes, sediment transport, and stratigraphy. Specific topics considered include the following:

a. Depth of closure and extent of sediment transport landward and seaward of this zone.

b. Processes that cause cross-shore movement of sediment.

c. Processes that cause net offshore and net onshore movement of sediment.

d. Amount and physical characteristics of beach material lost to the offshore.

e. Long-term fate of sediment that has moved offshore.

f. Relationship between depositional structures and flow processes.

g. Impact of episodic storms on sedimentation.

This literature review will help to define the current state of knowledge concerning cross-shore sediment transport on the inner shelf and sediment exchange between the beach and the inner shelf. Discussions will revolve around how sediment transport on the inner shelf is related to the equilibrium profile, depth of closure, sedimentation and stratigraphic characteristics of the inner shelf, and differences in sedimentation/stratigraphic patterns between fair-weather and storm conditions.

While this literature review does not comprehensively review all published material concerning inner shelf cross-shore sediment transport, it does provide reviews of some of the more important studies. In addition, comprehensive lists of inner shelf cross-shore sediment transport studies are included in the bibliography sections of the appendices.

Outline of Chapters

This literature review is divided into five chapters and three appendices. Chapter 2 discusses the equilibrium profile and depth of closure, and the importance of the geologic framework on inner shelf changes. Chapter 3 addresses several topics that verify the cross-shore transport of sediment on the inner shelf. These topics include mechanisms of inner shelf sediment transport, surf zone and inner shelf cross-shore sediment transport, beach-inner shelf sediment exchange, storm/fair-weather sediment transport, and storm sedimentation models. Chapter 4 concerns the sedimentation structures and stratigraphy of the inner shelf and includes topics such as inner shelf sedimentary features, inner shelf stratigraphy, cross-shore stratigraphic sequences, and storm-related stratigraphy. Chapter 5 summarizes some of the more important findings of this review.

Appendix A provides a glossary of useful terms. Appendix B is a bibliography of cross-shore sediment transport studies organized by topic. Appendix C is a bibliography of cross-shore sediment transport studies with respect to topic and region.

2 Inner Shelf Concepts

Introduction

This chapter reviews concepts that are crucial in determining the geologic aspects of inner shelf cross-shore sediment transport. These concepts include the equilibrium profile, the depth of closure, and the effect of the geological framework on the equilibrium profile and cross-shore sediment transport processes. These concepts are of concern to the engineering and scientific community primarily due to the unquantifiable amounts of sediment that are transported onshore and offshore of the inner shelf. Additional references concerning these topics can be found under individual reference lists entitled "Equilibrium Profile/Profile Adjustment References" and "Depth of Closure References" in Appendix B.

Equilibrium Profile

The equilibrium profile was first defined by Fenneman (1902), who stated "There is a profile of equilibrium which the water would ultimately impart, if allowed to carry its work to completion." Additional equilibrium profile studies include those by Cornaglia (1889); Ippen and Eagleson (1955); Eagleson, Glenne, and Dracup (1961); and Zenkovich (1967), who argued in terms of the null point hypothesis. This hypothesis states that shoreward increases in wave orbital asymmetry should be counterbalanced by shoreward increases in bed slope, thus creating an equilibrium profile.

Studies at Mission Bay, California, and the Danish North Sea Coast by Bruun (1954) found that the average of field profiles fits the relationship:

$$h = A\,y^{2/3} \tag{1}$$

where

h = water depth

$A =$ scaling parameter dependent on sediment characteristics

$y =$ distance offshore

The findings of this study are complemented by several laboratory studies including Rector (1954); Eagleson, Glenne, and Dracup (1963); Swart (1974); and Vellinga (1983).

A model concerning shoreline change in response to rising sea level (known as the Bruun Rule) was introduced by Bruun (1962). In this study, Bruun stated that the equilibrium profile described by Equation 1 would translate landward and upward while maintaining the original shape of the profile (Figure 2). Additional inner shelf equilibrium profile model studies include Inman and Bagnold (1963), Bailard (1981), and Bowen (1980). These models assume that the oscillatory motion of waves is the most important criterion in the development of the inner shelf equilibrium profile.

Dean (1977) stated that the equilibrium profile occurs when bed shear stress and the energy flux dissipation rate (function of wave energy density and group velocity) become equal everywhere over the profile. Dean (1983) further defined the equilibrium profile as "an idealization of conditions which occur in nature for particular sediment characteristics and steady wave conditions."

In proposing a model of destructive forces acting in the surf zone that would affect the equilibrium profile, Dean (1977) also reconsidered the equilibrium profile relationship by analyzing 504 beach profiles along the U.S. Atlantic and gulf shores (taken from Hayden et al. (1975)). Dean developed the following relationship:

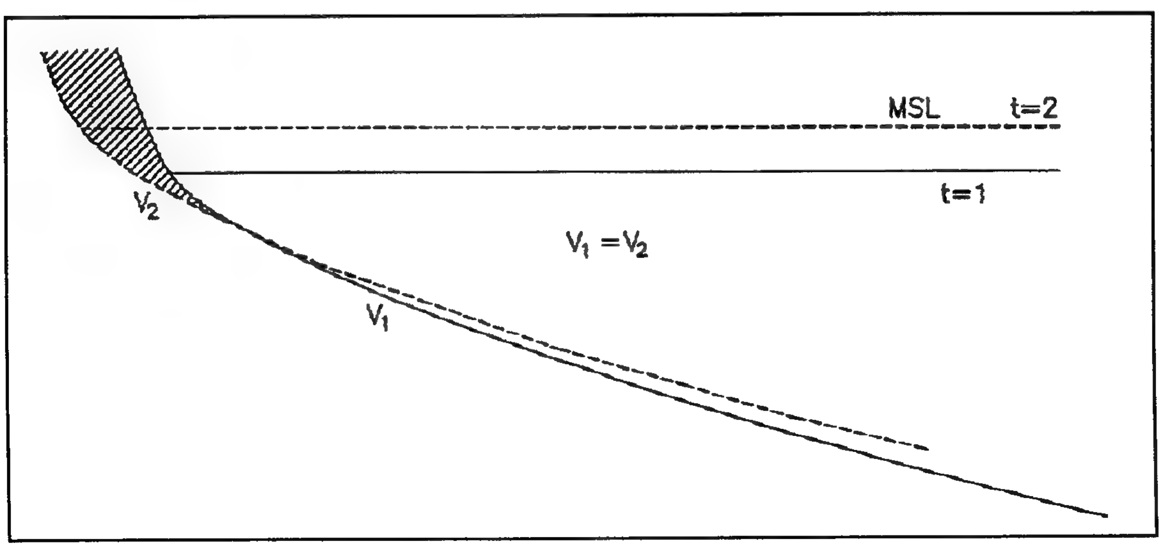

Figure 2. Translation of the original equilibrium profile in response to a rising sea level (after Bruun (1962))

$$h = A y^n \tag{2}$$

By applying the least squares fit to each of the profiles, Dean (1977) found ranges of the values for the parameters A and n (A ranged from 0.0025 to 6.31; n ranged from 0.1 to 1.4 with an average of 0.67, thus agreeing with Equation 1 of Bruun (1962)).

For Dean's (1977) model, he assigned a value of $n = 2/3$ when the rate of wave energy dissipation per *unit volume of the water column* is equal over the profile and $n = 2/5$ when the rate of wave energy dissipation *per unit area of the sea bed* is equal over the profile. Since the n value of 2/3 matched the average n for the 504 profiles (0.67), Dean (1977) stated that the critical factor in developing a profile of equilibrium must be the rate of wave energy dissipation per unit water column volume. Dean (1977) left the sediment scale parameter A as the only free variable. This resulted in a much smaller range of A values between 0.0 and 0.3.

Moore (1982), Dean (1987), and Kriebel, Kraus, and Larson (1991) related A to the sediment fall velocity using a single grain size for an entire profile.

Dean and Maurmeyer (1983) review several profile response models including those of Bruun (1962) and Edelman (1968, 1970), as well as several evaluations of Bruun's model including those of Schwartz (1965, 1967), Dubois (1975, 1976, 1977) and Rosen (1978). Dean and Maurmeyer found that:

a. Existing shore response models are useful for predicting long-term evolution due to relative sea-level rise. Better methods and field data are required to improve the capability of predicting depth of effective sand motion and the associated width of this zone.

b. The Bruun rule has been validated qualitatively and, to the limit of our knowledge of the relevant processes, quantitatively for the case of nonbarrier island systems. Dean and Maurmeyer (1983) state that for barrier island systems which migrate landward, their own model is more appropriate.

c. Of the existing models of the Bruun type, the Edelman (1970) model represents profile evolution as a continuing process and is therefore probably more representative of long-term response.

d. There is a need for application of improved profile response models that incorporate the effects of noncompatible sediment eroded and gradients in longshore sediment transport.

e. There is a need for improved definition of the detailed dynamics of beach profile response. This will probably require laboratory and field measurements under long-term and short-term (storm) events.

Larson (1991) described the profile of equilibrium as occurring when: "A beach of specific grain size, if exposed to constant forcing conditions, normally assumed to be short-period breaking waves, will develop a profile shape that displays no net change in time."

Dean (1991) listed four characteristics commonly associated with equilibrium beaches:

a. They are usually concave upwards.

b. The smaller the sand diameter, the more gradual the slope.

c. The beach face is usually planar.

d. Steeper waves result in more gradual slopes.

Pilkey et al. (1993) contend that the profile of equilibrium equation is inadequate to define the inner shelf profile shape and therefore should not be used as a basis for predictive models of profile evolution. First, although the equation provides an average inner shelf profile cross section, it does not effectively describe the true profile shape as it tends to ignore the effects of bars, and oversimplifies wave-inner shelf interactions. However, the equation does provide a useful guide particularly for long-term response of the "average profile." Secondly, the inner shelf is composed of various sediment grain sizes. The assignment of a value of 0.67 to the variable n in the profile equation, thus leaving a smaller range of values of the sediment scale parameter A (of 0.0 to 0.3), implies that beach profile shape can be calculated from sediment characteristics (particle size or fall velocity) alone.

Pilkey et al. (1993) state that the profile shape of the inner shelf is due to many factors, including the following:

a. Wave climate (particularly the frequency of big storms).

b. Sediment supply.

c. Rate of shoreline and inner shelf retreat.

d. Surficial sediment grain size.

e. Underlying geology (Figure 3).

Depth of Closure

The model proposed by Bruun (1962) concerning shoreline change in response to rising sea level also introduced the concept of *depth of closure* - "the point on the equilibrium profile beyond which there is no

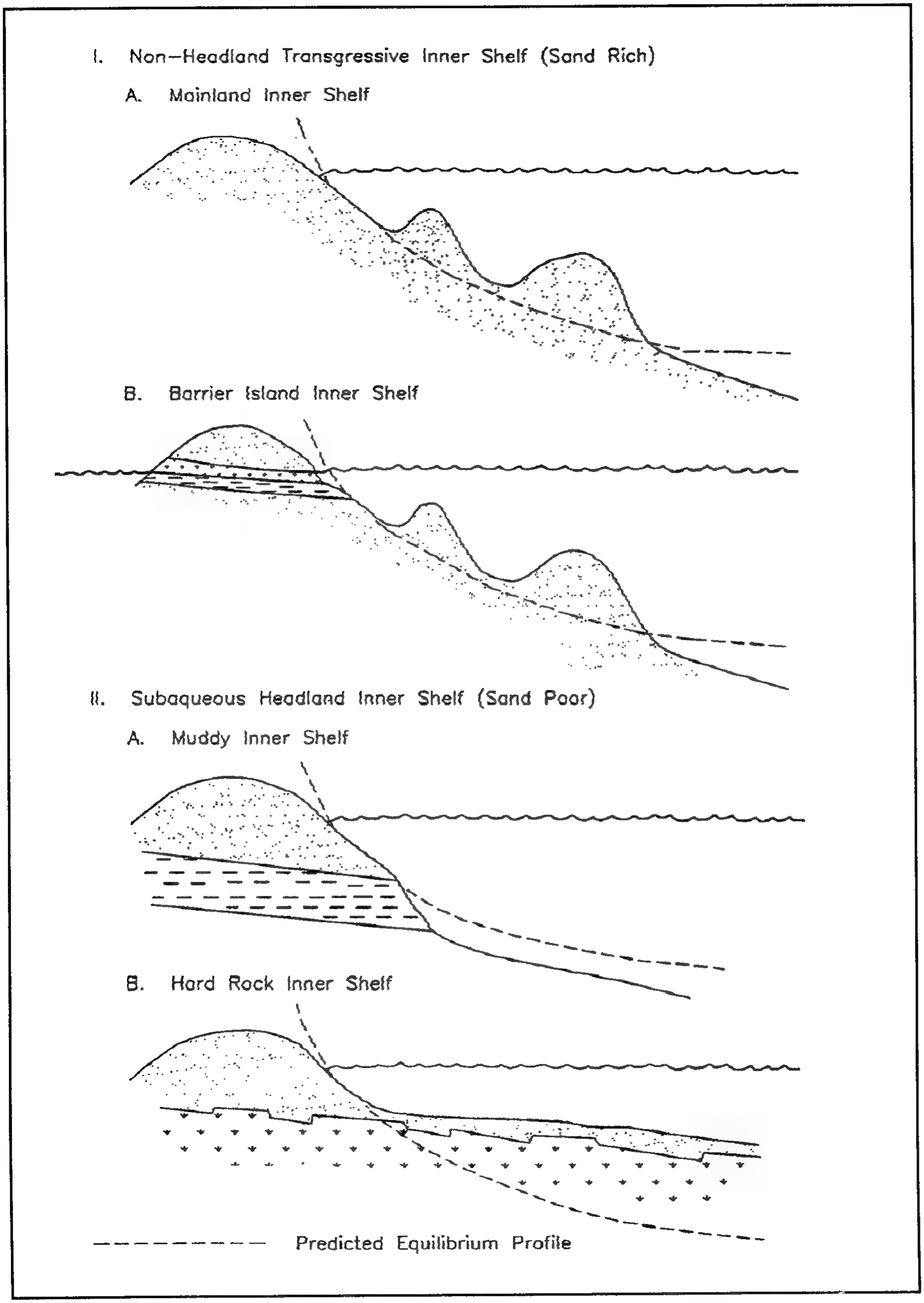

Figure 3. Possible inner shelf types resulting from different characteristics of underlying geology (after Pilkey et al. (1993))

significant net offshore transport of sand." Bruun examined evidence for the capability of offshore currents to transport sediment beyond the equilibrium profile closure depth. He chose 18 m as a "reasonable assumption" for this closure depth. He based this on the depth at which there is no measurable (within the error bars of profile measurement) change in pre- and post-storm inner shelf profiles.

Hallermeier (1978, 1981a) presented a model to estimate the seaward limit of sediment transport resulting from erosion (or offshore sediment transport). He developed a simple predictive equation, based on laboratory studies, to estimate the annual depth of the seaward limit. He defined two limits to an area he called the *shoal zone* (Figure 1). In the shoal zone, "surface waves are likely to cause little sand transport; ...waves have neither strong nor negligible effects on the sand bed" (Hallermeier 1981a). The seaward limit to the shoal zone (d_i) is the depth limit to sediment motion initiation by normal waves. This implies that significant onshore-offshore sediment transport is restricted to water depths less than d. The *offshore zone* is seaward of the shoal zone and is characterized by insignificant onshore-offshore transport by waves.

The landward limit of the shoal zone (d_1) separates the shoal zone and the *littoral zone*. The littoral zone is characterized by significant longshore and onshore-offshore sediment transport due to increased bed stress and sediment transport by breaking and near-breaking waves. According to Hallermeier (1977), d_1 can be described by a critical value of a sediment entrainment parameter (Φ_c) in the form of a Froude number:

$$\Phi_c = U_b / \gamma' g d = 0.03 \tag{3}$$

This critical value assumes that an intensely agitated bed usually exists seaward of the surf zone. Hallermeier (1977) suggested an analytical approximation, using linear wave theory for shoaling waves, to predict an annual value of d_1:

$$d_l = 2.28 H_e - 68.5 (H_e^2 / g T_e^2)$$

where

d_l = annual depth of closure below mean low water

H_e = nearshore nonbreaking wave height exceeding 12 hr/yr

T_e = corresponding wave period

g = acceleration due to gravity

According to the above equation, d_1 is primarily dependent on wave height with an adjustment for wave steepness.

Depth d_1 is considered the depth of closure and is used in estimating offshore closure limits for use in beach fill design. Hallermeier (1977) defined depth of closure as the minimum water depth at which no measurable or significant change in bottom depth occurs based on profile surveys.

To emphasize the importance of differences in wave and sand characteristics and wave variability on open sea coasts, Hallermeier (1978, 1981b) computed the depths d_1 and d for 30 sites on the Pacific, Atlantic, and Gulf of Mexico coasts using the wave climate study of Thompson (1977) and data from the Littoral Environmental Observation (LEO) Program. For the Gulf of Mexico coast (seven sites), the d_1 and d values were -4.2 m and -9.9 m, respectively. For the Atlantic coast (11 sites), d_1 and d were -5.7 and -22.1 m, respectively. D_1 and d values at the Pacific coast (12 sites) were -6.9 and -42.9 m, respectively. Differences in d_1 and d values stated above are a result of differences in significant wave height, wave period, and mean sediment grain diameter.

Boyd (1981) documented that the maximum depth of the initiation of sediment movement (similar to Hallermeier's (1981a) d) at the New South Wales, Australian continental shelf fluctuates with wave conditions (Figure 4). For instance, for wave height of 0.5 m and periods of 7 sec, this depth is -10 m; for wave height of 2 m and periods of 12 sec, this depth is -60 m.

Kraus and Harikai (1983) defined depth of closure as the minimum depth at which the standard deviation in depth change decreases markedly to a near constant value.

Birkemeier (1985) compared data from two profiles located in Duck, North Carolina, between August 1980 and December 1982 to Hallermeier's equation by measuring wave conditions that existed between profile surveys that exhibited offshore sand movement (Figure 5). Birkemeier (1985) found good agreement with the form of Equation 4, but recomputed the coefficient to better fit the data. He also found reasonable agreement using only H_e in Equation 5:

$$d_l = 1.57 H_e \tag{5}$$

where

d_l = nearshore limit, or closeout depth relative to mean low water

H_e = peak nearshore storm wave height, which is exceeded only 12 hr/year

He stated that this equation is probably site-specific.

Kraus (1992) conceptualized that the beach profile responds to wave action between two limits, one limit on the landward side where the wave runup ends and the other limit in deeper water where the waves can no

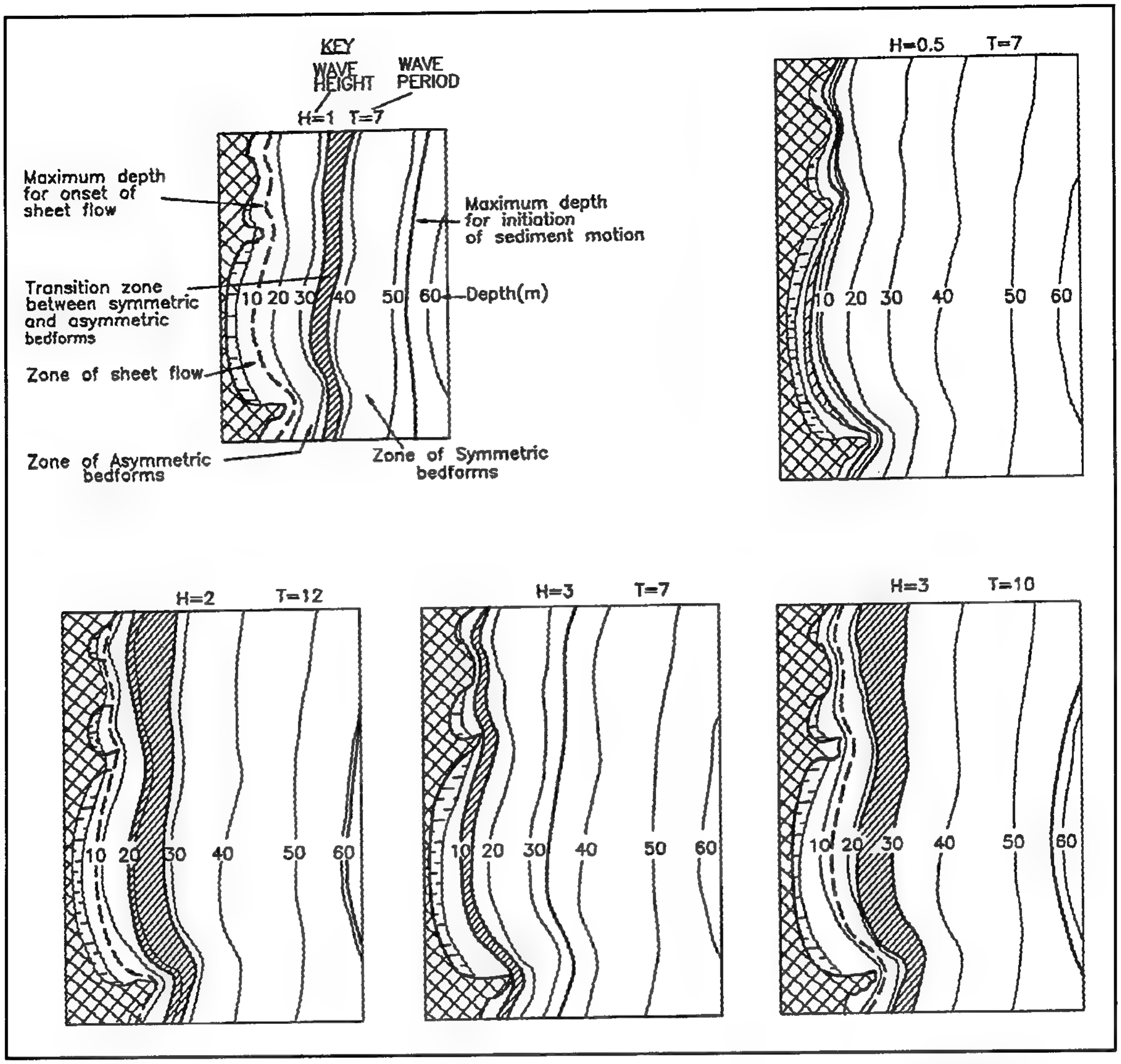

Figure 4. Fluctuations of inner shelf bed form zones and initiation of sediment motion with respect to significant wave Height (H) and period (T) (after Boyd (1981))

longer produce a measurable change in depth. He calls this latter limit, the minimum water depth at which no change occurs (as measured by engineering means) the depth of closure. The depth of closure is not the location where sediment ceases to move, but that location of minimum depth where profile surveys before and after a period of wave action, a storm perhaps, lie on top of one another.

Kraus (1992) also stated that the depth of closure is time-dependent, that is, dependent upon the transporting capacity of the particular incident waves. For example, we expect the average depth of closure for the summer to be less than that in winter. Similarly, the "storm of the decade" will alter the profile elevation to a much greater depth than occurs during a typical storm season. In engineering projects, the depth of closure is best determined through repeated accurate profile surveys, such as performed with a sled.

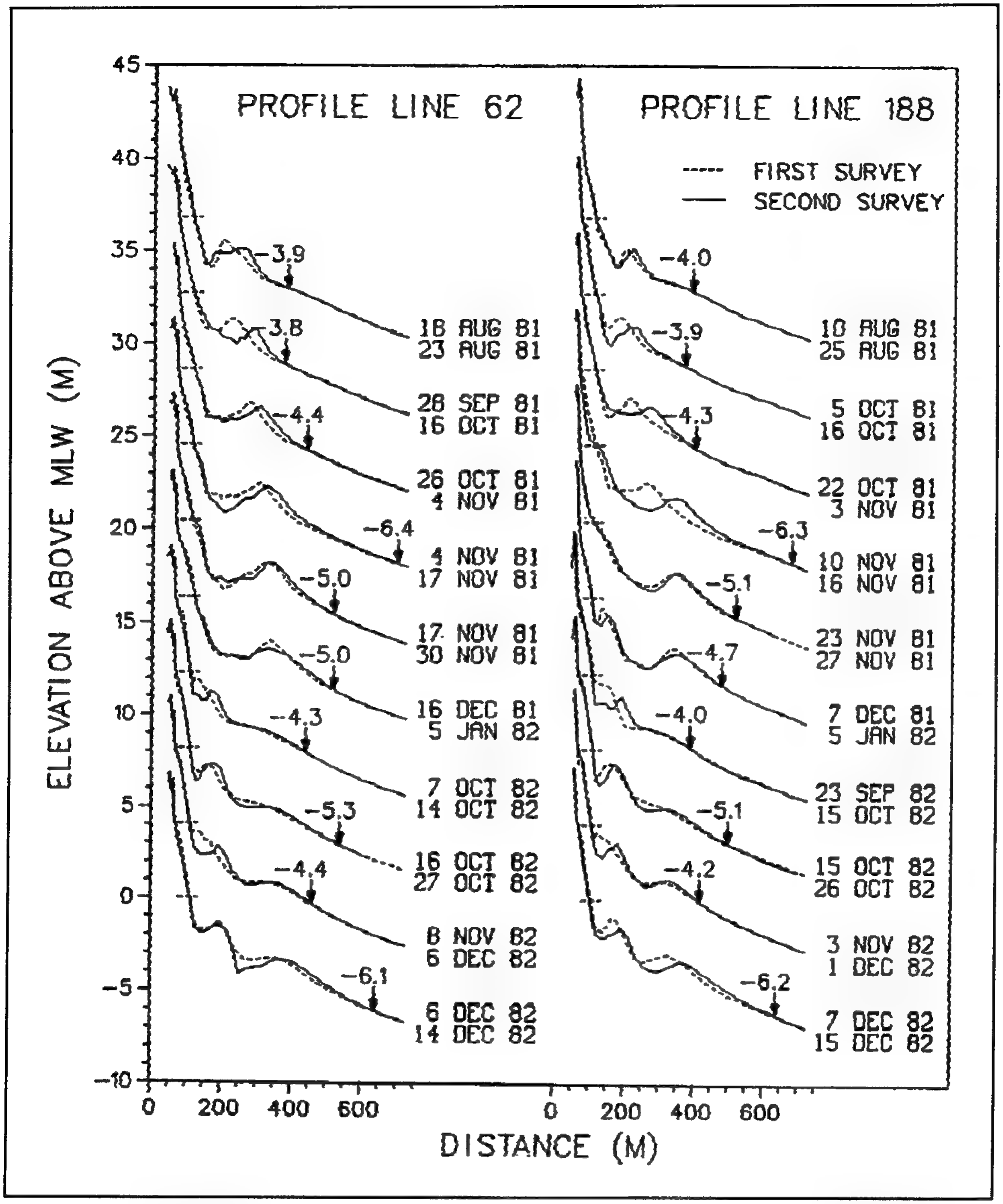

Figure 5. Survey data from Duck, NC, from August 1981 to December 1982 showing fluctuation of closure depth as indicated by vertical arrows (after Birkemeier (1985))

Pilkey et al. (1993) state that one of the most essential assumptions that must hold true for the concept of the equilibrium beach profile to be valid is: "There must exist a closure depth beyond which there is no net offshore or onshore transportation of sediment - a depth of no net sediment movement to and from the inner shelf even during storm-induced downwelling events." Pilkey et al. (1993) also defined the depth of closure as the depth where no vertical changes to the bed take place and where grain size distribution remains constant. Pilkey et al. (1993) state that the depth of closure does not exist, as field evidence shows that large volumes of sand may be moved beyond the closure depth. Such movement occurs mostly during offshore-directed storm flows. Studies in the Gulf of Mexico measured offshore bottom currents of up to 200 cm/sec and sediment transport to the edge of the continental shelf (Hayes 1967a,c; Morton 1981; Snedden, Nummedal, and Amos 1988). The amount of sediment moved offshore was large, but it was spread over such a large area that the change in seabed elevation could not be detected by standard profiling methods (Hayes 1967a,c) (±10 cm).

Several studies have found closure depths ranging from -5 m to -30 m for the U.S. Atlantic coast. Birkemeier (1985) stated that the measured depth of closure at Duck, North Carolina, fluctuates between -3.9 m and -6.4 m. However, the first conspicuous inner shelf configuration change at Duck occurs at -15 m, where sediments change from well-sorted fine sand to muddy fine sand with the fines bound in fecal pellets (Wright, in press). Perhaps this depth is more likely to be the maximum depth of normally occurring, shore-normal sediment exchange. This compares to Hallermeier's (1981b) seaward limit of sediment motion initiation (d) of -22.1 m for the Atlantic coast.

Depth of closure estimates using Hallermeier's (1977) method and the hindcast data of Jensen (1983) include Brevard County, Florida (-7.1 m); Walton County, Florida (-6.4 m); and Virginia Beach, Virginia (-5.5 m) (Hansen and Lillycrop 1988). Pearson and Riggs (1981) state that the depth of closure at Wrightsville Beach, North Carolina, was at least -16 m based on the presence of beach sediments at this depth. Wright (1987), in inner shelf studies including the use of bed elevation changes and sediment and profile data, shows that the depth of closure was located between depths of -10 and -30 m depending on regional energy regimes.

An additional estimate of depth of closure for the U.S. Atlantic coast is -9 m as presently used in engineering project design. This is the estimated depth where waves first affect the bottom as they move onshore, and there is no measurable (within the error bars of the profiling method) change in pre- and post-storm inner shelf profiles. In addition, sand ridges and irregular topography are typically located onshore of this closure depth while a uniform sloping shelf is located seaward.

Where Equation 4 predicts closure during the annual extreme 12-hr event, there exists a deficit of knowledge in predicting the depth of closure as a function of time. In order to develop a predictive method to

determine the time-dependent cross-shore transport of beach nourishment material, Stive et al. (1992) extend the annual shoreward boundary d_l (Hallermeier 1981b), by replacing the significant wave height exceeded 12 hr/yr (H_e in Hallermeier's 1977 equation) with the significant wave height exceeded 12 hr/return period (y) ($H_{e,y}$). Stive et al. (1992) considered an ideal model profile upon which a hypothetical beach nourishment was placed and subjected to the nearshore wave climate synthesis (function of H_{sig}) of Thompson and Harris (1972). They determined that d_l varied greatly during different wave conditions and return periods (Table 1). In addition, by assuming that beach nourishment volume decreases as a thinning wedge in the offshore direction, the spreading evolution and beach nourishment foot (depth of which beach nourishment migrates) may be approximated by applying the extension of the Hallermeier (1977) equation.

Table 1
Variation in Depths of D_l and Nourishment Foot for Different Wave Conditions and Storm Return Periods (from Stive et al. (1992))

Wave Steepness (H_i/L_o)	1/y Year Storm	D_l (Hallermeier 1981b) (m)	$D_{nourishment\ foot}$ (m)
0.01	1	7.3	8.1
	2	8.0	8.9
	5	8.9	9.9
	10	9.6	10.9
0.03	1	6.5	5.6
	2	7.2	6.1
	5	8.0	7.4
	10	8.7	8.6

Stauble et al. (1993) analyzed 3.5 years of profile data from Ocean City, Maryland, considering both storm and normal wave conditions. Twelve profile lines extended over 5.6 km of beach, and each consisted of seven or more surveys to the -9-m depth contour. Stauble et al. (1993) found that the depth of closure ranged between -5.5 m and -7.6 m, averaging -6 m. In addition, the profile at the northern end of the survey extent (103rd Street) was found to be steeper and without bars, while that of the southern end (37th Street) was shallow with bars. However, they suggest that more studies are required to relate the depth of closure to bar evolution.

Inner Shelf Geologic Framework Importance

Coastlines characterized by limited sand supplies, such as much of the U.S. Atlantic margin, are significantly influenced by the geologic framework occurring underneath and in front of the inner shelf (Figure 3). Passive margin coastlines, in particular, are significantly influenced by the geologic framework occurring underneath and in front of the inner shelf. This underlying geological framework can act as a subaqueous headland or hard ground that dictates the shape of the inner shelf profile and controls beach dynamics and the composition of the sediment.

The Atlantic coast of North America is an example of a coast affected by its geological framework. The advance of glaciers during the Pleistocene Epoch (characterized by continental glaciations at North America from approximately 2 million years to 10,000 years before present (ybp) (Evernden et al. 1964, Pratt and Schlee 1969) extended as far south on the Atlantic coast as northern New Jersey. North of the moraine terminus, glacial moraines composed of till (mixture of clay, silt, sand, gravel and boulders) underlie much of the land, islands (i.e. Long Island, Nantucket, and Martha's Vineyard), and offshore banks (i.e. Georges and Nova Scotian Banks). Coastal erosion of some of these features provides a variety of materials to the continental shelf. Conversely, south of the glacial moraine (Mid-Atlantic coast south of New Jersey), sediments are dominated by riverine sediments of piedmont streams that intersect the coastal plain strata.

Along the North Carolina coast, Pilkey et al. (1993) discuss that there exist three categories of underlying geologic framework which influence the inner shelf profile shape:

a. Subaerial headlands, which are composed of semi-indurated to indurated Pleistocene Epoch or older deposits incised by a wave-cut platform with a perched sand beach on the platform.

b. Submarine headlands, composed of semi-indurated to indurated Pleistocene Epoch or older units, which form the platform upon which the modern barrier island is perched and either crop out on the eroding inner shelf or occur on the inner shelf as paleotopographic highs in front of the modern inner shelf.

c. Nonheadland-transgressive inner shelf, commonly composed of Holocene Epoch (the Epoch from approximately 10,000 ybp to the present, which follows the continental glaciations of the Pleistocene Epoch) peat and mud deposits that extend from the modern estuaries, under the modern barrier islands, to crop out in the surf zone and inner shelf.

The Pleistocene section of the entire North Carolina coastal system represents a complex record of multiple cycles of coastal deposition and erosion in response to numerous glacial-eustatic, sea-level cycles (Riggs, Cleary, and Snyder, in press). During each glacial episode, fluvial channels severely dissected previously deposited coastal systems. The subsequent sea-level transgression then produced a ravinement surface that migrated landward and further eroded large portions of previously deposited coastal sediments by inner shelf erosion. This process of older units supplying sediment to the inner shelf of barrier islands was termed shoreface, or inner shelf, bypassing by Swift (1976). The fluvial channels were sequentially backfilled with fluvial, estuarine, and shelf sediments. Present day sea level has produced a modern sequence of coastal sediments that have been deposited unconformably over the eroded remnants of Pleistocene sequences composed of different lithofacies. Niedoroda, Swift, and Hopkins (1985) stated that this seaward thinning and fining veneer of modern inner shelf sediments over the older Pleistocene lithofacies is ephemeral and easily removed from the inner shelf during major storms.

On a smaller scale, the Nags Head/Kitty Hawk and the Rodanthe/Buxton areas on the Outer Banks of North Carolina, although separated by only 40 km, have distinctly different geological settings resulting in significantly different inner shelf profiles (Pearson 1979) (Figure 6). At the Nags Head/Kitty Hawk area, the inner shelf profiles contain two major sediment units including a modern inner shelf sediment wedge, composed primarily of reworked inner shelf sediments that thin in a seaward direction. These form a thin blanket over the in situ relict sediments that will ultimately crop out on the inner shelf. Pearson (1979) stated that this modern sediment wedge is periodically stripped away during extreme high-energy periods; thus exposing, possibly eroding, and transporting the relict units. By this mechanism, relict sediments are eroded and introduced into the modern sediment regime. In addition, the relict sediments underlying the thin, variable inner shelf sand sheet must also have a major impact upon the shape of the entire inner shelf profile.

In the Rodanthe/Buxton area, the inner shelf is controlled by Pleistocene hard-bottom topographic features that act as headlands and intersect the lower beach face at acute angles. These topographic features are believed to be a result of indurated Pleistocene stratigraphic units which outcrop in the Rodanthe area (Pilkey et al. 1993). These features include Wimble and Kinnakeet Shoals, permanent features up to 6 m in relief (Figure 7).

According to Pilkey et al. (1993), these vastly different inner shelf features have the following characteristics:

a. They dramatically affect the cross section of the inner shelf and beach profile.

b. They create major changes in the orientation of the barrier island (particularly at Rodanthe).

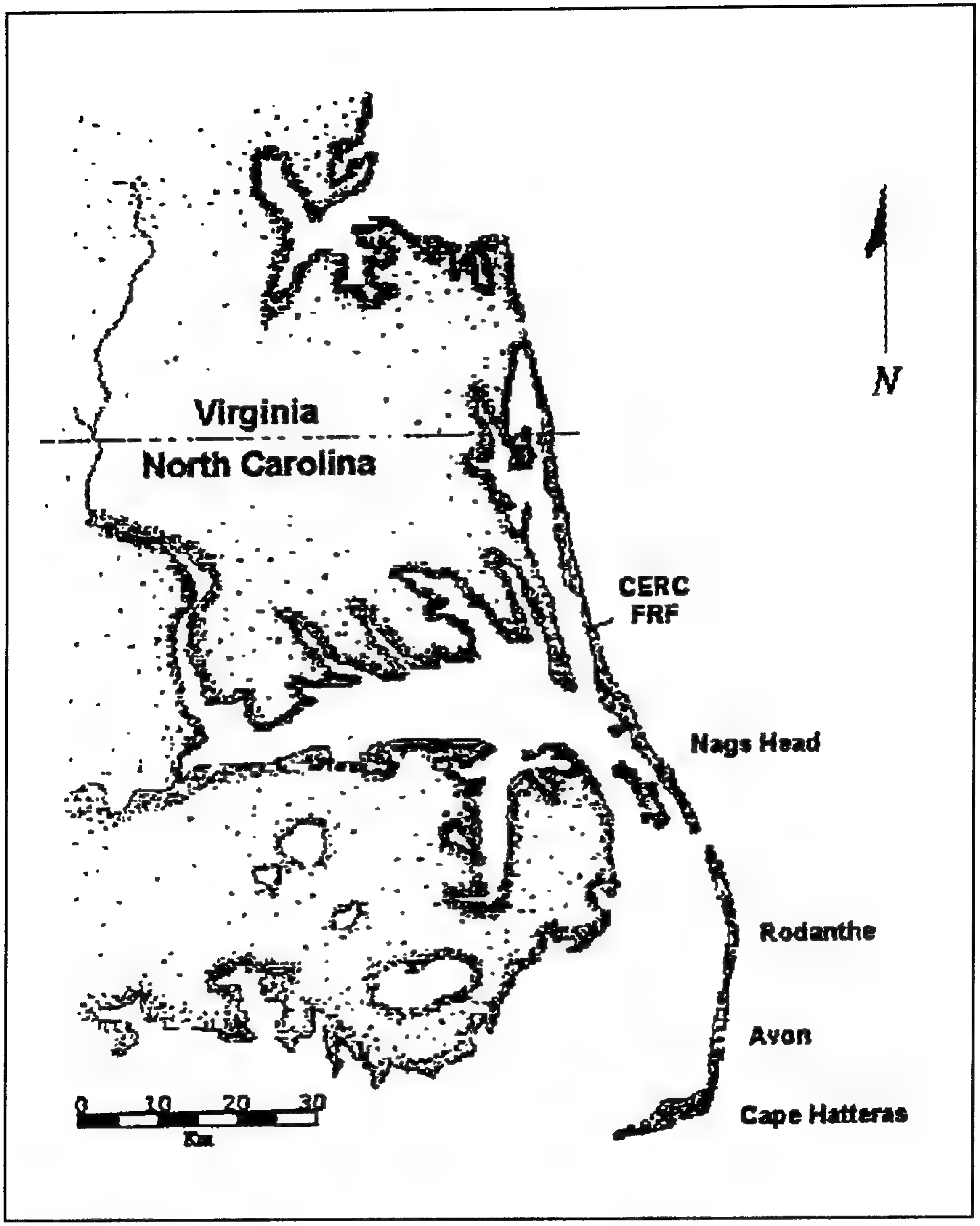

Figure 6. Location of the Outer Banks of North Carolina

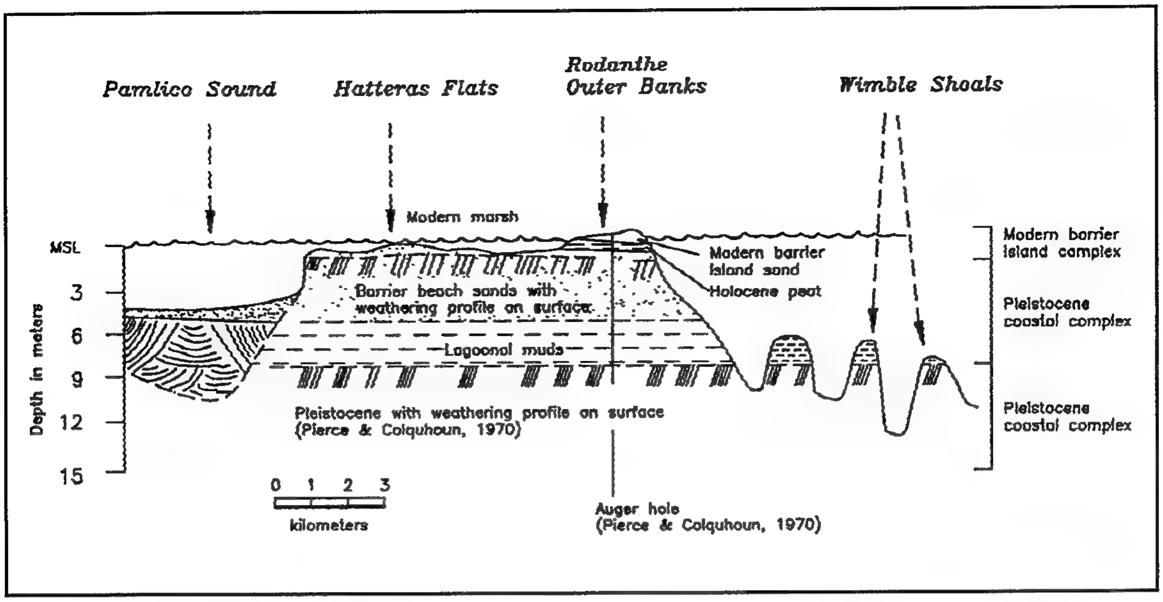

Figure 7. Geologic cross section through the Outer Banks at Rodanthe showing the Pleistocene units cropping out on the inner shelf forming Wimble Shoals (after Pilkey et al. (1993))

c. They are not in equilibrium with incoming wave energy, suggesting that these features erode.

d. They have dramatic impacts upon the energy regime affecting the adjacent inner shelf through wave refraction and setup.

In addition, the geomorphic nature of an area must also be considered when determining mechanisms and resulting shelf sediment transport. In examining patterns of sedimentation on the continental shelf, Swift (1976) examined the mechanisms by which the nearshore is penetrated (at the inner shelf/oceanic process boundary and at river mouths) and how sediment is injected into the shelf system. He found that the original mode of formation of the coast and surrounding areas had a large effect on present day sedimentation patterns. Swift (1976) differentiated between allochthonous and autochthonous settings. Allochthonous shelves (shelves presently composed of sediment formed elsewhere and subsequently deposited on the shelf) are typically floored by fine sands to muds (due to the introduction of riverine sediment through river-mouth bypassing) and are usually featureless, as these fine sediments travel in suspension. In addition, there is little bed form formation, as fine sediments have low angles of repose. Autochthonous shelves, or shelves presently composed of sediment originally derived from previous erosion of the shelf in its present location, are covered by coarser- grained sand of local origin.

3 Evidence of Cross-Shore Sediment Transport

Introduction

This chapter examines literature concerning evidences of cross-shore transport of sediment on the inner shelf. Patterns and mechanisms of sediment transport on the inner shelf, particularly in the cross-shore dimension, and of beach-shelf sediment interchange are poorly understood (Wright et al. 1991). Consequently, the generation of predictive theories which address these mechanisms and effectively recreate their effect on the cross-shore transport of sediment across the inner shelf is very difficult. Several authors (Wright 1987, Nummedal and Snedden 1987, Pilkey et al. 1993) concur that a model directly relating cross-shore sediment transport to transport mechanisms/processes is needed.

Additional topics discussed in this chapter include surf zone and inner shelf cross-shore transport of sediment, interchange of sediment between the beach and the inner shelf, and if this interchange results in the loss of sediment from the beach/inner shelf system to the outer shelf, storm/fair-weather sediment transport and storm sedimentation models. The purpose of the section concerning cross-shore sediment transport is not to provide a comprehensive review of all the theories of cross-shore sediment transport, but to discuss some of the evidences of this phenomenon and their relation to the theories of cross-shore sediment transport on the inner shelf.

Mechanisms of Inner Shelf Sediment Transport

The research of Wright et al. (1991) showed that bidirectional cross-shore sediment transport on the inner shelf is an exceedingly complex phenomenon driven primarily by shoaling waves, wind- and tide-generated currents, wave-current interactions, gravity-induced downslope transport, mean flows, and geostrophic circulation. However, these mechanisms have not been prioritized in terms of relative importance.

The mechanisms of cross-shore sediment transport are listed below and are more precisely documented in the literature by numerous authors as best summarized in part by Boyd (1981), Nummedal and Snedden (1987), Wright (1987), and Pilkey et al. (1993):

a. Waves and wave-driven currents, including:

(1) Powerful wave-orbital motions (Harms, Southard, and Walker 1982; Walker, Duke and Leckie 1983; Duke 1985; Duke 1987; Duke 1990) and resulting orbital asymmetry (Gilbert 1889; Wells 1967; Nielsen 1979; Hallermeier 1981a; Trowbridge and Madsen 1984; Swift and Niedorada 1985; Dean and Perlin 1986).

(2) Wave-induced upwelling and downwelling currents resulting from onshore/offshore movement of surface water and return bottom flows (Morton 1981, Snedden 1985, Wright et al. 1991).

(3) Wave-induced rip currents (Bowen and Inman 1969; Cook and Gorsline 1972; Reimnitz et al. 1976; Seymour 1983; Field and Roy 1984; Wright and Short 1984; Cowell 1986; and Wright et al. 1986).

(4) Sediment diffusion arising from gradients in wave energy dissipation associated with incoming incident waves (Wright et al. 1991).

(5) Sediment advection caused by wave orbital asymmetries associated with incoming incident waves (Wright et al. 1991).

(6) Long-period oscillations, which may be a more important process for cross-shore sediment transport in higher energy wave environments (Wright et al. 1991).

(7) Interactions between groupy incident waves (alternating high and low waves and forced long waves) (Shi and Larsen 1984, Dean and Perlin 1986, Wright et al. 1991).

(8) Groupy long waves (a forced long wave of infragravity frequency resulting in alternating high and low waves) (Shi and Larsen 1984, Dean and Perlin 1986, Wright 1987).

b. Wind- and tide-driven currents including:

(1) Semidiurnal and diurnal tidal currents (May 1979, Wright 1981).

(2) Strong, unidirectional currents from wind forcing (Morton 1981).

(3) Wind-induced upwelling and downwelling currents resulting from onshore/offshore movement of surface water and return bottom flows (Niedoroda et al. 1982; Morton 1981; Snedden 1985; Wright et al. 1986, 1991).

(4) Tidal currents.

(5) Storm surge ebb currents (Brenchley 1985).

c. Interaction of waves and currents (Butman, Noble, and Folger 1977; Lavelle et al. 1978; Grant and Madsen 1979a, 1986; Vincent, Young, and Swift 1982; Nielsen 1983; Shi and Larsen 1984; and Wright et al. 1991) including:

(1) Subharmonic and infragravity wave orbital interactions with the bottom sediment and with wave-induced longshore currents (Wright and Short 1984).

(2) Interactions between oscillatory flow and mean flow (Lundgren 1973; Smith 1977; Bakker and Van Doorn 1978; Grant and Madsen 1979b, 1986; Kemp and Simmons 1982; Wiberg and Smith 1983; Christofferson and Jonsson 1985; Coffey and Nielsen 1987).

d. Gravity-induced downslope transport often of highly concentrated sediment (Bruun 1962, Hayes 1967a, Dean 1977, Kobayashi 1982, Pilkey et al. 1993).

e. Forcing mean flows, which dominate and cause offshore transport during storms and contribute significantly to cross-shore sediment flux during fair-weather and moderate energy conditions (Wright et al. 1991).

f. Geostrophic circulation (Ekman spiral) and its superposition on wave motions (Komar 1976; Swift et al. 1983; Vincent, Young, and Swift 1983; Cacchione et al. 1984; Allen 1982; Neshyba 1987; Nottvedt and Kreisa 1987; Nummedal and Snedden 1987; Swift and Nummedal 1987).

g. Small-scale boundary layer processes (Wright 1994).

h. Physical oceanographic processes including oceanic currents (Csanady 1972; 1976; 1977 a,b; 1982; Csanady and Scott 1974; Halpern 1976; May 1979; Schwab et al. 1984).

Additional mechanisms contributing to cross-shore sediment transport include:

a. Storm surge-controlled breakout of coastal lagoons (Hayes 1967a, b, c), tidal inlets, and submarine canyons.

b. Turbidity currents (Bates 1953; Hayes 1967 a, b, c; Brenchley 1985; Seymour 1986; Wright et al. 1991).

c. Beach state (e.g. the first winter storm moving much more sediment than subsequent storms) including beach slope (Bascomb 1951, King 1972, Komar 1976, *Shore Protection Manual* 1984).

d. Formation of shell lags and a wide variety of bed forms (ranging from ripple marks to offshore bar systems)(Pilkey et al. 1993).

e. Organic scum layers (Pilkey et al.1993).

f. Variations in sediment pore pressure (Pilkey et al.1993).

g. Variations in the degree of sediment compaction and consolidation between storms (Pilkey et al. 1993).

h. Irregular inner shelf shapes (bedrock) which affect wave refraction patterns (Pilkey et al. 1993).

i. Coastal jets (Csanady 1972, 1977b; Csanady and Scott 1974; Ludwick 1977).

j. Topographic gyres (Bennet 1974, Csanady 1975).

k. Kelvin waves (Munk, Snodgrass, and Gilbert 1964; Munk, Snodgrass, and Wimbush 1970; LeBlond and Mysak 1977).

l. Vertical density stratification (Wright 1987).

Surf Zone Cross-Shore Sediment Transport

Much is known about nearshore sediment movement under shoaling waves (Komar 1976) and the documentation of cyclic patterns of surf-zone change (Wright et al. 1979, Nummedal and Snedden 1987). It has been documented that the most important concepts of surf zone dynamics and sediment transport are:

a. Orbital asymmetry (as expressed by second-and higher-order Stokes theory and supported by Gilbert (1889), Wells (1967), Hallermeier (1981a), Swift and Niedoroda (1985)).

b. Radiation stress theory and derived understandings (Longuet-Higgins and Stewart 1964).

c. Standing long waves and edge waves of infragravity frequency (Guza and Thornton 1985a).

Two useful models include Bailard's (1981) energetics model, which estimates sediment flux from measured wave and current data over the surf zone, and Guza and Thornton's (1985a, b) model, which is concerned with surf zone conditions where bed shear stresses and energy dissipation are strongly dominated by waves. Equations of both models help to determine if the cross-shore component of the immersed weight sediment transport within the surf zone is onshore or offshore.

A laboratory model developed by Hattori and Kawamata (1980), and its comparison with field data, is one approach which concerns the cross-shore transport of sediment in the surf zone. This model is based on the concept of the balance of power extended on sand grains generated by breaking waves, the beach slope, and the effect of gravity. Hattori and Kawamata theorized that cross-shore transport of sediment in the surf zone is a function of the dimensionless fall-time parameter as described by:

$$C = \frac{\left(\frac{H_o}{L_o}\right)\tan\beta}{W_s\left(\frac{d_{50}}{gT}\right)}$$

where:

$C =$ a constant determined from laboratory and field data

when

$C <$ 0.5 onshore transport results - accretive profile

$=$ 0.5 no net transport results - equilibrium profile

$>$ 0.5 offshore transport results - erosive profile

$tan\ \beta =$ bottom slope in the surf zone

$W_s =$ fall velocity of a sand grain of diameter d_{50}

$T =$ wave period

$H_o =$ deepwater significant wave height

$L_o =$ deepwater wavelength

Hattori and Kawamata (1980) continue that net cross-shore transport in the surf zone is a result of the stirring power P_s (which is a function of submerged weight of sand grains, maximum wave-induced velocity, bottom slope in the surf zone, water depth at the breaking position, and width of the surf zone) and the resisting power P_r (which is a function of fall velocity of a sand grain and the submerged weight of the sand grain. When $P_s > P_r$, sand grains keep in suspension due to breaking waves, and

sand grains are transported seaward in the form of a cloud by wave-induced currents (Sunamura 1980). When $P_r > P_s$, sand grains tend to roll and jump as bed load and move shoreward.

Inner Shelf Cross-Shore Sediment Transport

Introduction

Understanding of surf zone processes can be applied, at least in concept, to processes occurring on the inner shelf. For instance, Wright et al. (1991) applied surf zone sediment transport equations of Bailard (1981) and Guza and Thornton (1985 a,b) to predict inner shelf cross-shore sediment transport. Wright et al. (1991) found poor agreement between these surf zone and inner shelf sediment transport equations. Wright et al. (1991) state that these types of equations are needed to better predict cross-shore sediment transport on the inner shelf.

For wind-driven current patterns, Vincent, Young, and Swift (1983) divide the inner portion of the coastal ocean into the following three zones based on controlling sediment transport mechanisms:

a. *Geostrophic* (offshore; seaward of approximately the -15-m depth).

b. *Transition.*

c. *Friction-dominated* (seaward of the surf zone to approximately -10-m depth).

Landward of the 10-m contour in the friction-dominated zone, sediment transport rates are on the order of 1×10^{4} g/cm/sec and are primarily a function of asymmetric wave orbitals while seaward of the 10-m contour in the geostrophic zone, sediment transport rates are approximately 1×10^{2} g/cm/sec (Vincent, Young, and Swift 1983).

Geostrophic zone

Geostrophic circulation of ocean waters and sediment transport in this zone are controlled by the following factors:

a. *Cross-shore mean bottom currents* resulting from wind shear and tide-related currents.

b. *Currents generated by the Coriolis force.*

c. *Upwelling/downwelling conditions.*

The Coriolis force is defined as an apparent force resulting in the path deflection of an object due to the earth's rotation (Neshyba 1987). In the Northern Hemisphere, an object or water body undergoing movement on the earth's surface will be deflected to the right (clockwise) of the movement. The Ekman transport or drift, a function of the Coriolis force, states that as winds exert friction drag over an ocean of uniform density, a thin layer of surface water moves at an angle from the original wind (to the right in the Northern Hemisphere). This rotation continues as subsurface parcels of water are also rotated by the Ekman transport in that same direction. Therefore, there is a depth at which the water moves opposite to that of the surface wind (Neshyba 1987). Nummedal and Snedden (1987) have documented the Ekman transport in a three-layer inner shelf flow model, which shows that if surface currents are obliquely onshore, currents at mid-depths in the water column will be alongshore. Bottom currents will be oriented obliquely offshore.

Upwelling and downwelling currents are also geostrophically controlled currents that form due to orientation of the wind direction near a coast. For instance, upwelling conditions occur when offshore-directed winds transport surface waters in an offshore direction. Surface waters are then replaced by subsurface water and sediment, which moves onshore. Downwelling conditions, conversely, occur as onshore-directed winds transport the surface water onshore. Surface waters are then reflected by the beach, thus creating offshore-directed return flow of subsurface water parcels and sediment transport.

On the west coast of the United States, winds from the south will tend to deflect surface waters in a clockwise direction, or onshore, thus resulting in downwelling of deeper water parcels. Winds from the north will be deflected offshore, thus resulting in upwelling of deeper water parcels. On the east coast of the United States, upwelling tends to occur when winds are from the southwest, south, or northwest, while downwelling tends to occur when winds are from the northeast (Swift 1976).

Wright et al. (1986) conclude that northeaster storms create strong, southerly jet-like flows along the mid-Atlantic Bight. These flows affect the floor out to depths as far as -8 m, which results in downwelling and offshore sediment transport.

Friction-dominated zone

In the friction-dominated zone, a multitude of mechanisms affect inner shelf cross-shore sediment transport (see previous list of mechanisms of inner shelf cross-shore sediment transport). Overall, Wright et al. (1991) found that incoming incident waves were of primary importance in bed agitation (shear stress) and suspension of sediment on the inner shelf, while near-bottom tide- and wind-induced mean flows were of primary importance in the cross-shelf transport of sediment on the inner shelf. Wright et al. (1991) state that this mean-flow-generated cross-shore

transport of sediment was dominant or equal to that generated by incident waves in all cases and at all times.

Pilkey and Field (1972) and Wright et al. (1991) distinguish between the primary causes of onshore and offshore cross-shelf sediment transport. Pilkey and Field (1972) summarize the mechanisms of onshore transport of sediment on the inner shelf, which include wave and tidal current phenomena such as:

a. Onshore component of asymmetrical wave orbitals under shoaling conditions.

b. Onshore-oriented dominating tidal flood currents in shallow water.

c. Both the onshore and offshore components associated with storm-induced bottom currents.

In addition, Wright et al. (1991) state that incident waves are an important mechanism of the onshore transport of sediment.

Sediment transport mechanisms documented to cause onshore and offshore cross-shore sediment transport include the following:

a. Orbital asymmetry.

b. Interaction of incident waves with infragravity waves and mean offshore flows.

c. Wave groupiness.

d. Slope of the shelf and effects of gravity.

e. Rip currents (Wright et al. 1991).

Discussion of these mechanisms of inner shelf offshore and onshore cross-shore transport follow.

Orbital asymmetry. Findings by Cook and Gorsline (1972) during studies at Palos Verde, California, as supported by May (1979) and Wright et al. (1991) indicate that orbital asymmetry-created currents during wave shoaling transport sediment in both the onshore and offshore directions. These findings include the following:

a. Both onshore and offshore asymmetry of currents were documented during wave shoaling. Long-period swells and offshore breezes cause a net onshore transport of sediment, while short-period waves and onshore winds are associated with neutral or offshore flow.

b. Swell characteristics also affect water drift, in that long-period waves have onshore pulses which prevail temporarily, and thus cause net onshore transport of sediment.

c. Tidal surge asymmetry includes components of both onshore and offshore sediment transport across the inner shelf.

d. Tidal flux does not have a significant effect on surge asymmetry. However, May (1979) found that 35 percent of the kinetic energy of currents above the 30-m isobath in the Northern Middle Atlantic Bight was at a tidal frequency, thus indicating the importance of tidal currents in affecting sediment transport on the shelf. In macrotidal environments tidal currents probably dominate the inner shelf transport (Wright 1981).

e. Wind affects the ratio for durations of current flow and bottom drift, thus resulting in upwelling and downwelling flow.

Cook and Gorsline (1972) and Trowbridge and Madsen (1984) discuss the importance of sediment transport under asymmetric waves and related orbital asymmetry in generating both onshore and offshore components of cross-shore sediment transport. Also, time and space variations in bed roughness when considering orbital asymmetry can affect both magnitude and direction of sediment transport. Oscillatory currents over rippled beds can cause a significant phase angle between instantaneous suspended sediment concentration and instantaneous velocity, resulting in sediment flux in a direction opposite to the net current or wave-induced mass transport (e.g. Nielsen (1979)).

Larsen (1982) also found that the net offshore transport of sediment on the inner shelf is a function of the net offshore orbital asymmetry of waves. Currents forced by the radiation stress of variable amplitude swell (the higher waves suspending the sediments) are an important mechanism in suspending sediments resulting in the cross-shore transport of sediment on mid-continental shelves.

Smith and Hopkins (1972) found that orbital asymmetry-created currents during wave shoaling are the dominant control of net onshore transport of sediment, primarily of coarse material, on the inner shelf.

Wave-current interaction. Grant and Madsen (1979a, 1986) theoretically discussed combined wave-current bottom fluid shear stress and stated that the actual transport across the inner shelf is, in most cases, the result of wave-current interaction. Effects of wave-current interaction on the boundary layer include the following:

a. Increases in rate of frictional dissipation of waves.

b. Reduction in mean current speed near the bed.

c. Increases in bottom shear stress due to a combination of components.

The importance of wave-current interaction in determining the magnitude and direction of sediment transport is also considered by Vincent, Young, and Swift (1983). They found that when wave orbital velocities and slowly varying bottom boundary layer velocities are combined, stronger onshore combined flow results. Moreover, depending on bed roughness and the horizontal angle between wave incidence and the mean current, the vector resultant of the sediment flux may be opposite that of the mean current.

Wave groupiness. Wave groupiness is also an important factor of net offshore transport of sediment across the inner shelf. Wave groupiness causes space and time variations in wave amplitude and in radiation stress (S_{xx}). Thus, momentum balance requires that slowly time-varying mean water level (η_f) be depressed and elevated under high and low waves, respectively (where S_{xx} is greater and less, respectively). Variances in f cause a long-period infragravity wave. This infragravity wave has peaks at low primary waves which result in onshore sediment transport (i.e. shoreward values of f (or the cross-shore long wave flow constituent)) and troughs at high primary waves, which result in offshore sediment transport (i.e. seaward values of f). Since the large primary waves in the trough of the long wave suspend more sand (offshore-directed) than the small primary waves of the long wave crest, there is a net seaward transport (Wright et al. 1991).

Gravity-induced currents. Gravity-induced inner shelf offshore-directed sediment transport (as stated by early references considering the equilibrium profile concept (e.g. Cornaglia 1889, Ippen and Eagleson 1955, Bruun 1962, Inman and Bagnold 1963) occurs due to the slope of the inner shelf being oriented in an offshore direction. This gravity-induced offshore transport of sediment is accentuated where fine-grained sediments are present, since these types of sediment can be easily suspended, especially during storm events.

Seymour (1986), in studying different models of turbidity currents and their relation to inner shelf transport, confirms that these currents transport nearshore sand in an offshore direction during storms.

Wright et al. (1991) noted that gravity plays a significant role during high-energy events when bed shear stress and suspended sediment concentration were greatest. If a density current develops, and the sediment is suspended at a greater rate than it is deposited, an autosuspending offshore-directed turbidity current can form. Kobayashi (1982), who developed a model for net downslope sediment transport by oscillatory flows acting on a gentle slope, found that gravity-induced offshore-directed transport of sediment is significant.

Rip currents. Rip currents are also important in transporting sediment in an offshore direction (Field and Roy 1984). Bowen and Inman (1969) and Cook and Gorsline (1972) report that during the winter season, cross-shore movement of sediment by rip currents is in an offshore direction. Once transported offshore, sediment is confined by predominant seaward oscillations caused by steep waves and strong winds. During summer, long-period swells transport sediment landward to replenish the beach. Cook and Gorsline (1972) also present a sediment transport system whereby sediment is transported offshore outside of the breaker zone by rip currents and general diffusion, and then onshore by wave action, which separates silt and clay from sand. Sand is then moved alongshore to depths dependent upon wave characteristics. Silt and clay are separated in the sorting process and move out of the coastal drift system in suspension.

Reimnitz et al. (1976) used side-scan sonar to show seaward-trending ripples out to depths of 30 m that are attributed to storm rip currents. Cowell (1986) measured rip currents off headland-bounded beaches during storms and measured velocities of greater than 1 m/sec extended to hundreds of meters past the surf zone. However, Field and Roy (1984) believe that rip currents probably do not transport sand to a depth greater than 45 m.

Seymour (1983), in experiments at Santa Barbara, Torrey Pines, and Virginia Beach (as part of the Nearshore Sediment Transport Study), also documented rip currents as a mechanism of offshore sediment transport. During periods of intense storm waves, Seymour (1983) documented the formation of offshore bars, particularly at Santa Barbara. The formation of these bars is attributed to excessive longshore sediment transport and rip current outlets during these storms.

Hyperpycnal plumes. Hyperpycnal plumes, or sediment/water flows of dense concentration that plunge under flows of less dense concentration associated with gravity flows (Bates 1953), may also result in seaward transport where fine-grained sediments are present (no autosuspension is needed). In studies by Wright et al. (1991), where bed slope was 0.6 deg, suspended sediment concentrations were as high as 10 g/l, and underflows were as thick as 2 m with downslope speeds of 10-40 cm/sec, Wright et al. (1991) attributed this offshore-directed sediment flow to a rise of 0.6 m in mean water level (during this particular storm) and a resultant strong seaward-directed downwelling flow.

Bar formation/migration. Osborne and Greenwood (submitted, 1992) determined that cross-shore sediment transport at a non-barred inner shelf in Nova Scotia and a barred inner shelf at Georgian Bay are similar and a function of the following parameters:

a. Local wind-forced low-frequency waves.

b. *Mean current flows* (in the Nova Scotia non-barred example, these flows were offshore-directed undertows).

Additional causes of non-barred inner shelf sediment transport include swell, while additional causes of barred inner shelf sediment transport include high-frequency wind wave oscillatory currents.

Osborne and Greenwood (1992) also differentiate between sediment transport at different locations on the bar. On the lakeward slope of the bar, a net offshore sediment transport component of mean currents results from the offshore flow of undertow and group-forced bound long waves, and the landward flow mechanism of wind wave oscillatory currents. This is in contrast to studies on Padre Island, Texas, by Hill and Hunter (1976) who show that net onshore bottom currents are dominant on the seaward side of the bars and the bar crests under normal breaking wave conditions of 0.3 to 1.0 m. On the bar crest, Osborne and Greenwood (1992) state that there was no net transport of sediment due to a balance between offshore mean transport (undertow) and onshore net oscillatory transport (interaction between both high- and low-frequency waves). Landward of the bar crest and in the trough, although the wind waves decrease due to dissipation of wave energy, suspended sediment transport by low-frequency waves is most important, thus transporting sediment in a predominantly onshore direction (Osborne and Greenwood 1992).

Sediment trends

Wright (in press), in a study at the Field Research Facility at Duck, North Carolina, documented that the grain size of the inner shelf over the upper 18 m exhibits a slight tendency to fine seaward (Figure 8). Fine to very fine sand (D_{50} = 0.09-0.13 mm) prevails, while silts and clays comprise 10-15 percent of the surficial sediment. This seaward-fining sequence is a result of decreases in energy in an offshore direction.

Different magnitudes and properties of offshore versus onshore flow across the inner shelf have resulted in the differential transport of fine versus coarse sediments. Smith and Hopkins (1972) state that during storm events fine material is transported offshore, while coarse material is transported onshore. They documented that fine sand moves as suspended load from the nearshore and is transported offshore during severe storms. During non-storm periods, both fine and coarse sand move onshore by wave-driven bottom currents, which have a net onshore component.

Basically, coarse material moves onshore due to the greater energy exerted by the onshore-directed wave orbitals which are shorter, and exert great velocities on the bed. Fine material moves offshore as suspended load by the offshore-oriented orbitals, which are longer and of less energy. Thus, the coarse material is moved onshore while the fine material moves offshore (Wright et al. 1991).

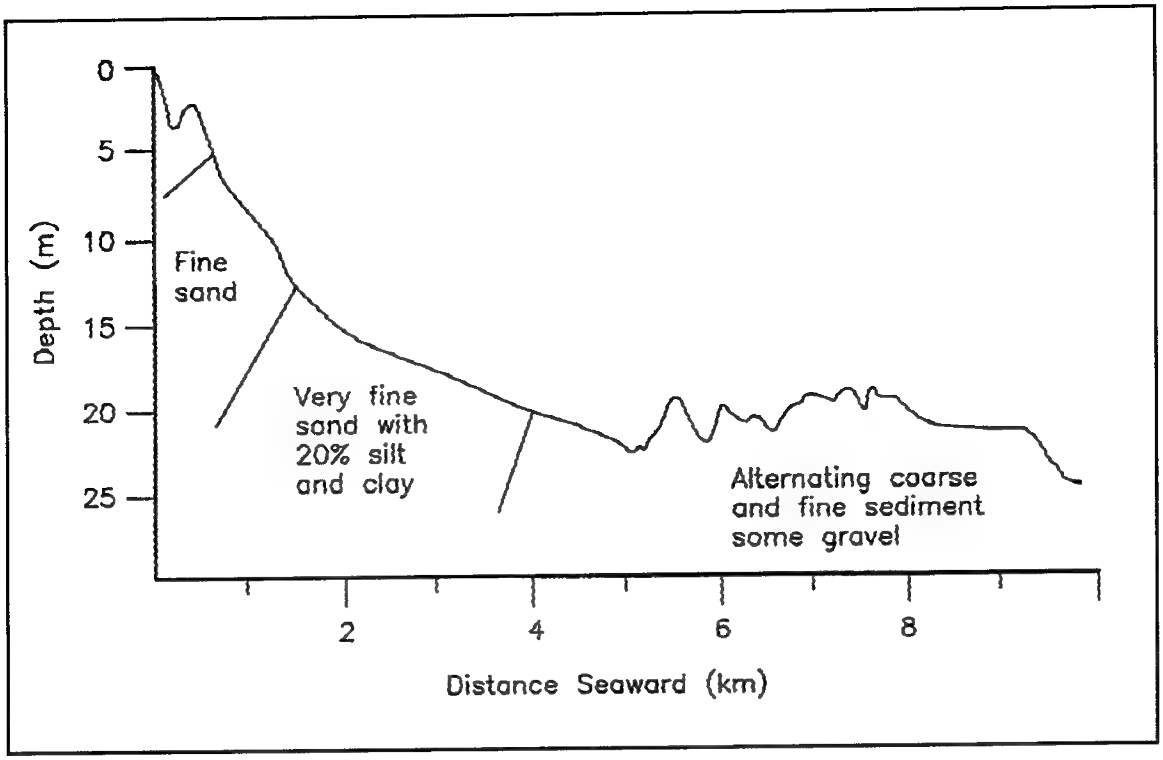

Figure 8. Cross-shelf profile of the inner shelf off Duck, North Carolina (after Wright et al. (in press))

Smith and Hopkins (1972) determined in their study of Columbia River sediments that an average particle on the shelf moves about 40 km/year in a longshore direction and 7 km/year in an offshore direction. The majority of this transport occurs only during a few storms each winter. Estimates of sediment transport indicate that the sand fraction moves much more slowly as bed load than the silt fraction as suspended load.

Seasonal effects on inner shelf cross-shore sediment transport

Seasonal cross-shore transport of sediment along the southern California coast has been documented by Shepard (1950), Shepard and Inman (1950), Inman (1953), Inman and Rusnak (1956), and Aubrey (1979). During summer, the subaerial beach accretes, while the offshore loses sediment. In winter, the subaerial beach erodes, while the offshore accretes. These changes are a result of variation in wave frequency and directional properties (e.g. Pawka et al. (1976)). Small-amplitude, long-period waves dominate in summer, while higher-energy, high-frequency storm waves dominate in winter.

Aubrey (1979) examines temporal changes in beach/inner shelf profile configuration using eigenfunction analysis of profile data for southern California profiles for a 5-year period. Two seasonal pivotal points separating eroding and accreting regions are documented at -2 m to -3 m, and at -6 m. A simple model of depth-dependent seasonal sand movement

shows that during initial winter storms, sand is eroded from both the foreshore and from depths of -6 m to -10 m and is deposited at the -2-m to -6-m water depth. During less energetic periods, sediment migrates both onshore (to the beachface) as well as offshore (to a depth of -10 m) from its winter site of deposition (-2 m to -6 m). This depth-dependent motion contradicts the single pivotal-point model previously suggested for nearshore seasonal cross-shore sediment motion and emphasizes the complexity of nearshore sediment transport. A sediment budget for seasonal cross-shore transport, based on the dual pivotal point model, consists of exchanges of 85 m^3/m at the -3-m pivotal point, and 15 m^3/m at the -6-m pivotal point. On a longer (5-year) time scale, beaches showed no erosion or accretion, suggesting that the limited coastal region is stable over this time period.

Beach-Inner Shelf Sediment Exchange/Losses

Now that evidence has been presented concerning the onshore and offshore components of cross-shore sediment transport, the actual exchange of sediment between the inner shelf and the beach is considered. Boyd (1981) emphasized that cross-shore sediment exchange represents a major contribution to the inner shelf sediment budget.

Studies by Pearson and Riggs (1981) extensively documented the exchange of sediment between the beach and the inner shelf at Wrightsville Beach, North Carolina. It is this study which has accentuated the importance of the permanent loss of sediment from the beach-inner shelf system. Two findings associated with this study are important. First, Pearson and Riggs (1981) observed the offshore transport of replenishment sand from Wrightsville Beach to a depth of -16.6 m. This is based on the presence of beach nourishment sand (fine to coarse-grained gray to black sand with oyster shells) which is easily distinguishable from North Carolina continental shelf sands, which are brown in color. This suggests that the depth of closure at Wrightsville Beach is at least -16.6 m.

Secondly, Pearson and Riggs (1981) state that periodic renourishment totalling 7,300,000 cu m of material placed since 1939 (which would cover a 23.3-km^2 area with a 14.6-cm layer of sediment) is being effectively and permanently removed from the nearshore system. This renourishment sand requirement has not decreased over time, indicating that the profile is not establishing an equilibrium profile. Pilkey et al. (1993) contend that if the concept of the equilibrium profile were valid, then the volume of sand needed to nourish the profile should decrease over the years as it accumulates above closure depth on the inner shelf.

In studies of Hurricanes Carla and Allen, and tropical storm Delia on the Texas shelf, Nummedal and Snedden (1987) document the cross-shore exchange of sediment as a great loss of sediment from the beach-inner shelf. They found that sand is moved offshore during storms due to

downwelling (three-layer flow) but is not returned onshore. Niedoroda, Swift, and Hopkins (1985) also supported the loss of sediment from the beach-inner shelf system only during storms. However, they state that some of the sand transferred from beach to inner shelf during storms will return.

Luternauer and Pilkey (1967) employ the use of minerals (i.e. phosphorite) at the North Carolina coast at Onslow Bay to document the interchange of sediment between the beach and the inner shelf. They found that the shelf is an important source of beach sediments. This suggests that the shelf is a major contributor of phosphorite to landward beaches. Another interesting finding of this study was that a small amount of longshore transport occurs on the shelf as phosphorite content is limited to Onslow Bay and does not spill over to other embayments. This indicates that phosphorite is a useful tool for determining sediment provenance and transportation.

Thus, several studies support the interchange of sediment between the beach and the inner shelf. However, there are examples in the literature where no sediment interchange occurs. For instance, Meisburger (1989) investigated the interchange of sediment between the beach and Gilbert Shoal, a nearshore linear shoal off Florida. He determined that the major sediment source to the beach is from littoral processes, while a lesser amount of sediment comes from the shoal. However, the shoal and surrounding seafloor receive little, if any, sediment from the beach or nearby St. Lucie Inlet. The shoal obtains sediment from the nearby shelf floor and from in situ shell production.

Depth of inner shelf sediment transport

When considering sediment interchange between the shelf and the beach, the next logical question is to what depth is sediment transported and/or affected on the continental shelf. This topic was previously considered in the "Depth of Closure" section of Chapter 2 as discussed by Draper (1967); Harlett (1972); Komar, Neudeck, and Kulm (1972); Smith and Hopkins (1972); Sternberg and Larsen (1976); Channon and Hamilton (1976); Sternberg and McManus (1972); Gadd, LaVelle, and Swift (1978); Vincent, Swift, and Hillard (1981); Larsen et al. (1981); and Wright et al. (1986). In addition, Grant and Madsen (1979a,b, 1986), Madsen and Grant (1976), Larsen et al. (1981), and Niedoroda et al. (1982) compute bed load transport at depths. Evidence of sediment transport at considerable depths (greater than -40 m) follows.

Direct current measurements on the central and outer continental shelf of Washington and Oregon by Smith and Hopkins (1972) at the -50-m and -80-m water depths showed that significant sediment transport in an offshore direction, most importantly by suspended load, occurs only during storms. A storm with current speeds of up to 60 cm/sec transports on the order of 6 m^3/hr/m of sediment of shelf length, while a 70-cm/sec storm

transports $15m^3$/hr/m of sediment of shelf length. Net transport of sediment is offshore. These data suggest that a single severe storm may be more effective in transporting sediment than several small storms.

Komar, Neudeck, and Kulm (1972) discuss the production of orbitals by surface waves, which in turn create ripples, and rework shelf sediments. Table 2 shows relationships between depth of rippling and a variety of surface wave conditions (after Komar, Neudeck and Kulm (1972)).

Table 2
Relationships Between Depth of Rippling and a Variety of Surface Wave Conditions (after Komar, Neudeck, and Kulm (1972))

Surface Wave Conditions	Wave Period (sec)	Significant Wave Height (m)	Depth of Rippling (m)	Orbital Diameter (cm)	Ripple Wavelength (cm)
Average Summer Waves	12	2.13	85	38.2	12.6
Average Winter Waves	12	3.05	99	38.2	12.6
Large Storm Conditions	12	9.14	138	38.2	12.6
Long-Period Storm Waves	15	9.14	204	47.7	10.3

Symmetrical (wave-generated) oscillatory shore-parallel ripple marks (see section in Chapter 4 titled "Examples of Inner Shelf Sedimentary Features" for additional information on ripple symmetry) exist on the Oregon continental shelf out to water depths of -204 m, while asymmetrical ripples are rare. Symmetrical ripples are covered by bottom orbital velocities (as calculated by the Airy wave theory) as well as unidirectional currents while asymmetrical ripples are believed to be produced by internal waves (15- to 30-min period), as they are more similar to unidirectional currents. It is believed that upwelling currents could not have formed ripples (Komar, Neudeck, and Kulm 1972).

Larsen et al. (1981) determined that at the -100-m depth on the Washington shelf, for sediment sizes 0.03-0.07 mm, a bottom oscillating current of 13 cm/sec is needed to suspend sediments. These types of currents and waves are common during winter storms in Washington, where 100-cm/sec velocities associated with 15-sec waves have been measured. Draper (1967) calculated that fine sand on the shelf edge of Britain would be moved at a depth of 183 m 20 percent of the year. Sternberg and Larsen (1976) found that relatively frequent grain motion occurs at the -75-m depth on the Washington shelf.

In addition, computations of bed-load transport by Madsen and Grant (1976) have shown that for conditions with 1.5-m, 13-sec waves, bed load was entrained to a depth of -16 m.

Storm/Fair-Weather Sediment Transport

Several researchers (Hayes 1967a,c; Murray 1970; Morton 1981; Green at al. 1988; Wright et al. 1991) have documented the differences in cross-shore inner shelf mechanisms and resulting sediment transport during fair-weather and storm conditions (refer to "Significant (Storm) Event References" in Appendix B for additional references concerning this topic).

Green at al. (1988) and Wright et al. (1991) in Mid-Atlantic Bight experiments measured suspended sediment movement, wave heights, and mean current flows between the -7-m and -17-m depth contours at Duck, NC, in 1985 and 1987 and at Sandbridge,VA, in 1988. The purpose of this work was to identify modes, directions, rates, and causes of shore-normal sediment flux over the inner shelf in response to different energy conditions (Table 3). Field measurements were compared to energetics mathematical models of sand transport (Bowen 1980; Bailard 1981; Guza and Thornton 1985 a,b; Roelvink and Stive 1989) who compared the contributions of mean and oscillatory flows, and separated cross-shore components of immersed weight sediment transport into bed load and suspended load.

Table 3
Summary of Environmental Conditions at Duck, North Carolina, for Different Events (after Wright et al. (1991))

Parameter	Summer Fair Weather July, 1987	Post-Hurricane Fair Weather August, 1991	Winter Swell-Dominated January, 1988	Extra-Tropical Storm October, 1991
Bed roughness	Large ripples, biogenic activity	Ripples on mounds and holes	Small ripples, irregular	Highly mobile plane bed
Depth of instrumentation	8 m	8 m	7 m	13 m
Current speed	9.0-16.5 cm/sec	10.6-13.0 cm/sec	4.0-13.6 cm/sec	2.0-49.5 cm/sec
Wave height	0.35-0.40 m	0.29-0.40 m	0.9-1.4 m	1.0 - > 4.0 m
Wave period	8.7-9.0 sec	7.0 sec	9.7-12.9 sec	9.0-14.0 sec
Wave/current angle	45°-75°	26°-34°	29°-66°	36°-85°

Fair-weather sediment transport

In documenting fair-weather processes, Green et al. (1988) and Wright et al. (1991) examined data collected at the -8-m and -17-m depths during two data collection periods at Duck (1985 and 1987). Green et al. (1988) and Wright et al. (1991) found that although tides and oscillatory wave motion strongly influence both onshore and offshore sediment transport

processes, mean cross-shore flows were of greatest importance. There existed no relation between bed stress by instantaneous cross-shore velocity and suspended sediment concentration. The mean cross-shore flow reversed with the tide. During high tide, weak offshore flows occurred, while during low tides stronger onshore flows resulted. Bed load and suspended load quantities were nearly equivalent.

In fair-weather conditions, Wright et al. (1991) found that cross-shore flows differed according to depth. Overall, flows at the -8-m depth tended to be more energetic and had greater sediment transport rates by an order of magnitude. At a depth of -8 m, suspended sediment transport, which was dominated by mean cross-shore flows, was predominantly offshore. However, these flows reversed direction more often than those at a depth of -17 m. Conversely, at a depth of -17 m, a slight landward flow from mean flow and oscillatory currents resulted.

Larsen (1982) stated that offshore sediment transport on the shelf is a slow but steady seaward motion of resuspended sediments. This contradicted the conclusions of other researchers (e.g. Wright et al. 1991) who stated that offshore sediment transport on the shelf occurred during a few events with a strong offshore component. The time required to establish steady flow conditions is approximately a tidal cycle offshore, but decreases to several hours at shallower depth at the inner shelf due to friction.

Moderate energy sediment transport

Moderate energy processes, and related sediment transport, as studied at Sandbridge, VA, in 1988, were dominated primarily by mean flows, incident wave orbitals, and tidal currents (Wright et al. 1991). The dominant flow was oriented onshore (which may be a function of tidal currents and upwelling from west winds during the study period). As in fair-weather processes, there was little relationship between suspended sediment transport and bed stress during moderate energy conditions. Suspended sediment concentration, which at times equaled 1.5 kg/m^3, varied considerably over the period. Tidal variation also occurred, as it did during fair-weather processes. However, in deference to fair-weather process periods, weak onshore currents occurred during higher tides.

Swell-dominated processes

Swell-dominated processes, as measured at Duck, North Carolina, in 1988 (during wave conditions of H_s of 0.85-1.4 m and periods of 10-14 sec), resulted in overall onshore flow (Wright et al. 1991). However, many flow reversals occurred due to constant weak offshore-directed cross-shore mean flows, which opposed high-frequency landward-directed wave-induced oscillatory flows. These wave orbital velocities (maximum of 0.5 m/sec) were the main source of bed shear stress.

Overall, during swell-dominated conditions, the bed was strongly agitated at all times (suspended sediment concentration exceeded 1.0 kg/m^3). Findings indicated that the suspended sediment load is dominant over bed load, and was directed onshore due to the landward-oriented incident wave orbital motion.

Storm-dominated processes

Storm-dominated processes were measured during a 'northeaster' storm at Duck, North Carolina, in 1985 (storm surge of 0.6 m; wave heights of 1-1.4 m, wave periods averaging 8 sec)(Wright et al. 1991). Sediment transport prior to the storm was bidirectional but was net offshore during the storm and was greater than that of fair-weather and moderate processes by an order of one to two magnitudes. This net offshore transport of sediment occurred due to onshore winds, the resulting 0.6-m rise in mean water level, and associated downwelling and offshore-directed bottom mean flows. However, this offshore sediment transport is much less than alongshore transport of sediment.

During storm-dominated processes, suspended sediment concentrations averaged above 1.0 kg/m^3 throughout the study and were up to 4.0 kg/m^3 associated with wave orbital velocities up to 1.0 m/sec (Wright et al. 1991). During the height of the storm, suspended sediment concentrations were 4,000 mg/l at 14 cm above the bed; 1,400 mg/l at 34 cm above the bed; and 200mg/l at 106 cm above the bed. Although there was a relationship between suspended sediment concentration and wave orbital velocity, there was no relationship between suspended sediment concentration and bed shear stress. The effect of the bed shear stresses on the bed (in order of occurrence) included:

a. Negligible changes in bed level response to the initial impulses of the storm including wind, mean and oscillatory currents, and suspended sediment concentration maxima.

b. Gradual, but significant, scour of the bed of 5 cm during the storm phase that followed the initial impulse.

c. Initiation of accretion of the bed during the second and stronger peak of the storm.

d. Rapid accretion of the bed (15 cm) during the waning phases of the storm (this accretion, the authors note, may be a migrating bed form or offshore pulse-like migration of sediment).

These bed level changes are believed to be associated with high-energy wind waves, which cause mixing and mobility of the upper sediment column thus causing offshore-oriented sediment exchanges (Wright et al. 1991).

Hayes (1967c) studied Hurricanes Carla and Cindy in the Gulf of Mexico to examine the direct effects of storm processes and sediment transport. They recorded cross-shelf thicknesses and textures of Hurricane Carla beds to a depth of -35 m off Padre Island, Texas, along 50 km of coast. Hayes (1967c) documented that sediment was transferred between the beach and the inner shelf in both the onshore and offshore directions. Before and during Hurricane Carla, mollusk shells, coral blocks, and other materials were transported onshore from water depths between 15 m and 25 m and deposited on the beach. Storm surge seaward-directed turbidity currents carried the sediment offshore. After the storm passed, offshore-directed currents associated with hurricane-generated channels deposited a 1.25-cm to 3.75-cm layer of sand over preexisting mud out to depths of -18 m. In addition, a graded layer of fine sand silt and clay (known as a turbidite) was deposited.

Summary

Green et al. (1988) document sediment transport changes according to different phases of the storm. During fair-weather conditions, although the waves were asymmetric in an onshore direction, the reversing tidal currents and resulting mean flow controlled inner shelf sediment transport. During the early phase of the storm, sediment transport was controlled by wind-driven jet-like flow (mean flow) with an offshore component. During the progression and towards the end of the storm, the waves were more organized and highly skewed in a onshore direction, thus enabling the highly skewed wave-orbital velocities to transport sediment in an onshore direction against the mean flow. Storm flow was dominated by suspended load, which accounted for 75 percent of the sediment volume.

In summarizing the findings of Green et al. (1988) and Wright et al. (1991), mean flows, interpreted to be related to tides, were dominant over incident waves in generating cross-shore sediment fluxes across the inner shelf. Cross-shore mean flows during fair-weather conditions were negligible, while these flows were greater than 20 cm/sec during storm conditions. Oscillatory flows associated with waves were 10 cm/sec and 100 cm/sec during fair-weather and storm conditions, respectively. Suspended sediment concentrations 10 cm above the bed were less than 0.1 kg/m^3 and 1-2 kg/m^3 during fair-weather and storm conditions, respectively.

Storm Sedimentation Models

Modeling of storm sedimentation is limited to the models of Dott and Bourgeois (1982); Walker (1984); Brenchley (1985); Duke (1985); and Duke, Arnott, and Cheel (1991), who base their models on the following parameters:

a. Textures in modern storm sediments.

b. Geostrophic flow concepts.

c. Results of flume experiments.

d. Inferred storm-generated structures within ancient sandstones to construct cross-shelf facies sequences dependent upon water depth, sediment availability, and storm parameters such as return frequency and strength.

Keen and Slingerland (1993a) note that while these models represent an important conceptual advance, they are qualitative and have not been tested against oceanographic data collected for that purpose, or compared to results of numerical experiments.

Keen and Slingerland (1993b) have constructed a three-dimensional numerical prediction model to hindcast the oceanographic and sedimentologic responses of the western Gulf of Mexico to four historical tropical cyclones.

The simulations of the numerical model by Keen and Slingerland (1993b) indicate that:

a. Onshore flow to the right of the storm track generally transports fine sediment landward.

b. Offshore flow to the left of the storm track transports coarser sediments seaward.

c. A right-to-left (facing the coast) alongshore flow transports finer sediment in deep water and coarser sediment in shallower water.

The models of Keen and Slingerland (1993a,b) suggest that coastal geometry is the controlling factor in determining sedimentation patterns, while in situ sediments are the main source of sediments to the inner shelf. Along the coast in front of each storm, the volume of sediment transported obliquely in a cross-shore direction is a function of the shelf gradient and coastal configuration. Steeper gradients constrain flow to a more longshore pattern. Concave coastlines promote greater shoreface erosion because of increased setup.

4 Sedimentary Features/ Stratigraphy of the Inner Shelf

Introduction

Mechanisms of cross-shore sediment transport on the inner shelf greatly affect sedimentary features including morphological signatures such as surficial bed forms, and stratigraphy (internal structure) of the inner shelf. The first studies of inner shelf sedimentary features and stratigraphy characteristics were those of Agassiz (1888), Grabau (1913), and Johnson (1919). Johnson (1919), who developed the first model of continental shelf sedimentary characteristics, stated that:

a. The shelf is a system in dynamic equilibrium both in terms of slope and grain parameters.

b. Given a nearshore sediment source, grain size decreases in an offshore direction due to decreasing wave energy.

Shepard (1932) stated that the shelf was composed of a mosaic of sediment sizes and types rather than a uniform seaward-fining trend in grain size. He suggested that these sediments were deposited during periods of lower sea level, particularly during the Pleistocene Epoch. Emery (1952, 1968) presented a classification of shelf sediments on a genetic basis considering the following types of materials:

a. Authigenic, or formed or generated in place (e.g. glauconite or phosphorite).

b. Organic, or relating to a compound containing carbon as an essential component (e.g. foraminifera, shells).

c. Residual, or relating to an accumulation of rock debris formed by weathering which remains in place (e.g. residual clay).

d. *Relict*, or remnant from an earlier environment such as a beach or dune.

e. *Detrital material*, or presently supplied from rivers, coastal erosion, and eolian or glacial activity.

Emery (1952) stated that in most coastal environments, the nearshore zone is composed of modern detrital sediments, while the shelf is composed of relict sands.

Curray (1964) stated that stratigraphy of the continental shelf is a function of the following:

a. Fluctuations in sea level.

b. Rate of sediment input to the continental shelf.

c. Sediment grain size and mineralogy.

d. Rate of energy input.

e. Rate of relative sea level change.

f. Continental shelf slope.

Curray (1964) found that the onshore (transgression)/offshore (regression) migration of the shoreline, and subsequent sediment dispersal and rate of net deposition/erosion of sediment on the continental shelf are functions of the rate of sea level rise (subsidence of the land) or sea level fall (emergence of the land) (Figure 9). Migrations of the shoreline and deposition of sediment on the continental shelf are important in understanding the paleogeography, sources, environments, and deposition mechanisms of sediments.

Examples of Inner Shelf Sedimentary Features

There exist a wide range of sedimentary features on the inner shelf ranging in scale from linear shoals (also known as ridge and swale topography) (hundreds of meters) to individual bed forms (centimeters to meters).

Large-scale sedimentary features

The large-scale sedimentary morphology of the middle Atlantic Bight was first extensively documented during the Inner Continental Shelf Sediment and Structure Program (ICONS) undertaken by the U.S. Army Corps

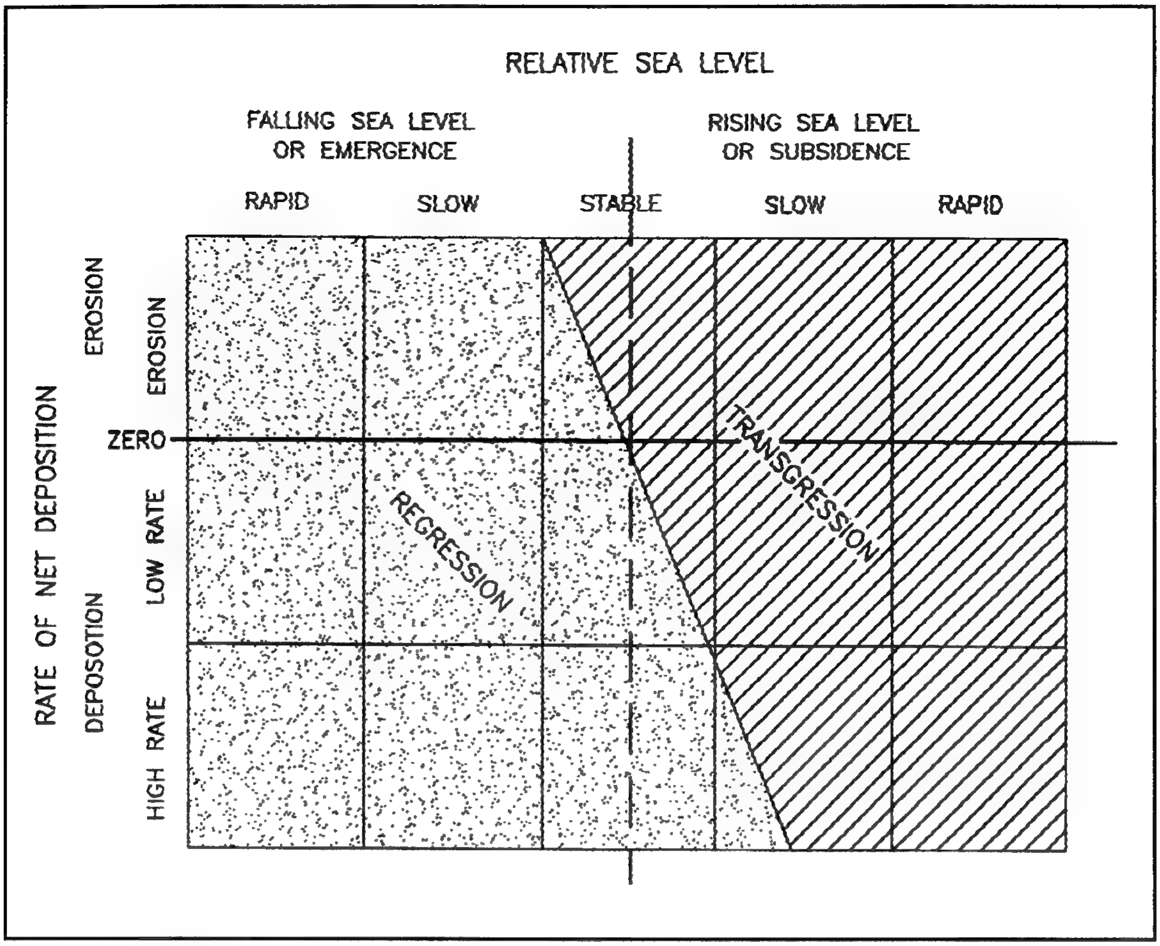

Figure 9. Relationship between rate of net sediment deposition/erosion and rate of sea level rise/fall (after Curray (1964))

of Engineers in the mid-1960s. This program was undertaken to accomplish the following:

a. Identify continental shelf sand bodies for beach nourishment purposes.

b. Garner a greater understanding of shelf sedimentation as it pertains to the supply of sand for beaches.

c. Increase understanding of changes in coastal and shelf morphology, longshore sediment transport, inlet migration and stabilization, and navigation.

d. Increase understanding of the geologic history of the continental shelf.

Additional studies of the Middle Atlantic Bight of North America include Veatch and Smith (1939), Shepard (1963), Emery (1966), Uchupi (1968), and Duane et al. (1972).

ICONS helped to identify the larger framework of geomorphic sedimentary features on the Middle Atlantic Bight of North America, including the following (Figure 10):

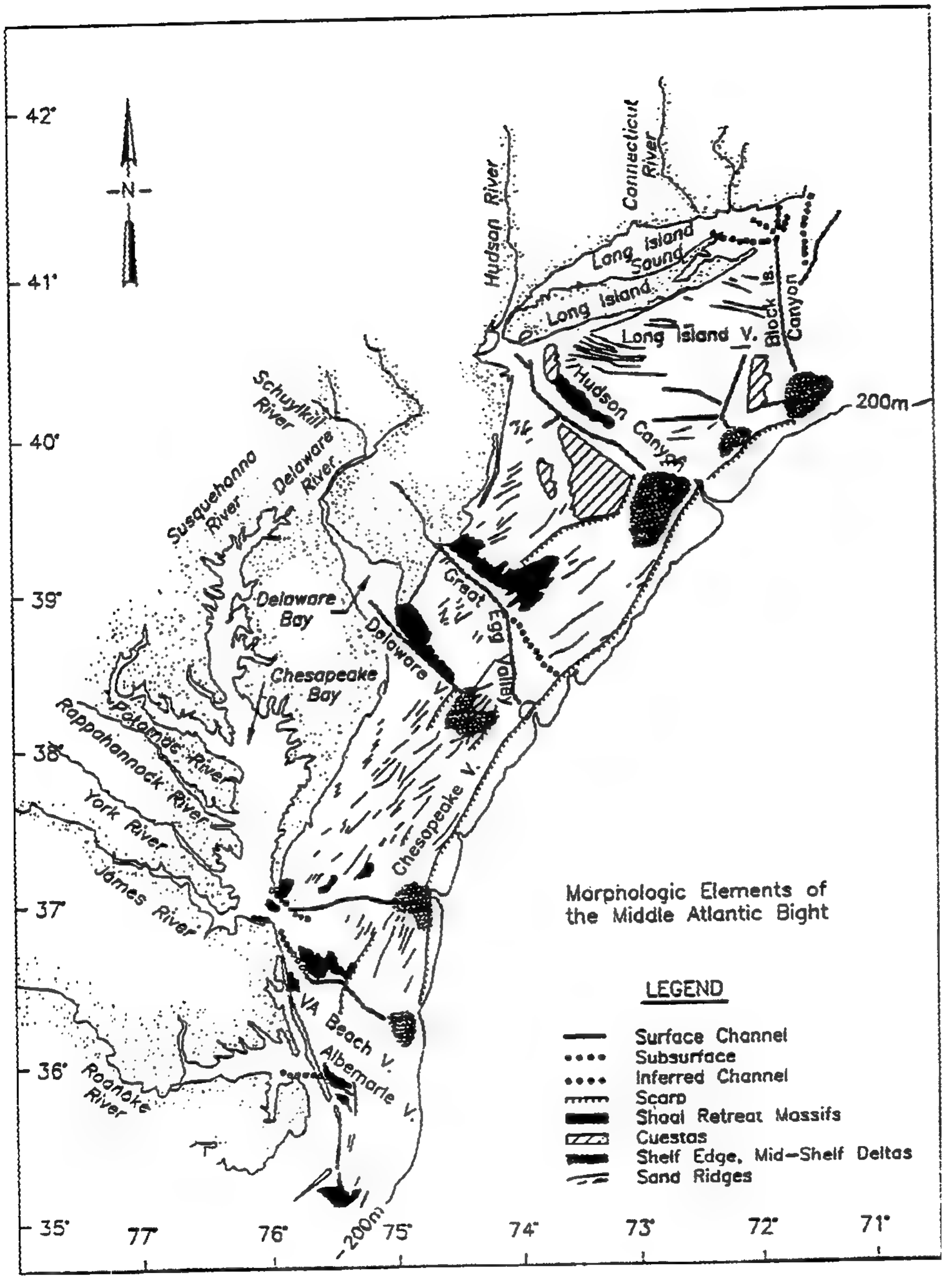

Figure 10. Morphology of the Middle Atlantic Bight (after Swift (1975))

a. Broad, flat plateaus.

b. Fluvial valleys and related deltas excavated during the Quaternary Period (from approximately 2 million ybp to the Recent (present) Period inclusive of the Pleistocene and Holocene Epochs)(Evernden et al. 1964, Pratt and Schlee 1969).

c. Shoal and retreat massifs (landward migration of deltas during transgression [or rising sea level]).

d. Terraces and scarps.

e. Cuestas.

f. Sand ridges.

Duane et al. (1972) summarized these studies and discussed both inner shelf-detached and shelf-attached shoals. Linear northeast-trending inner shelf-detached shoals trend from the shoreline at an angle between 5 deg and 25 deg, are located in water depths of up to -30 m, measure approximately 25 to 500 m in length, have reliefs of up to 10 m, have side slopes of a few degrees, and extend for tens of kilometers. These sand bodies are composed of well-sorted medium- to coarse-grained sands and are similar in lithology to adjacent beaches. In some instances, clusters of shoals merge with the shoreline in depths as low as 3 m.

Inner shelf-attached shoals are shoals that are landward of the wave base (about -8 m)(Duane et al. 1972)(although these features are located in the nearshore zone, they are not similar in nature to surf zone/nearshore bars). These shoals appear to form in response to the interaction of south-trending, shore-parallel, wind-generated currents with wave and storm-generated bottom currents during winter storms. Aggradation of crests occurs during storm waves, while degradation occurs during fair-weather waves. These shoals are believed to have formed during lower sea levels associated with the Wisconsin stage of glaciation (the most recent and farthest south continental glaciation advancement approximately 21,500 ybp to 10,000 ybp during the Pleistocene Epoch) (Evernden et al. 1964, Pratt and Schlee 1969). The shoals are modified by present-day coastal processes, as they are in equilibrium with shelf processes. If these shoals were not in equilibrium with present-day processes, they would erode and disappear.

Field and Roy (1984) also document elongate, shore-parallel shoals on the lower inner shelf in southeast Australia. These bodies are 10-30 m thick and parallel the coast for 40 km. The upper parts of these sand bodies are composed of sand transported downslope from the upper inner shelf and surf zone. Surface sediments of ridges are well-sorted and coarser than surrounding sediments. No seaward fining trend exists. Internally, beds are parallel to the slope of the inner shelf and there is no evidence of cross- bedding, thus making it difficult to determine the exact

seaward sediment transport mechanisms responsible for the formation of these structures. Field and Roy (1984) indicate that the most plausible mechanism is the seaward transport of sediment during storm-induced downwelling currents.

Cacchione et al. (1984), in a study associated with the Coastal Ocean Dynamics Experiment, have identified three types of sedimentary features of the Central California inner shelf up to 2 km from the coast in -65 m of water. These included:

a. Rocky outcrops.

b. Elongate depressions of low relief on the inner shelf slightly oblique or normal to the general trend of the isobaths. These depressions contain ripples (heights of 0.40 m; wavelengths of 1.7 m) believed to be formed by large-amplitude, long-period winter surface waves.

c. Smooth areas of no perceptible relief, but covered with well-defined wave ripples (heights of 0.02-0.05 m, wavelengths of 0.20-0.30 m).

The proposed generation mechanism of these features is storm-generated bottom currents associated with strong, storm-driven downwelling flows during late fall and winter, steered by underwater rock ledges which scour the surficial fine-grained sediment and expose the coarser-sand substrate in the depressions (Cacchione et al. 1984).

Small-scale sedimentary features

Bed form classification. Harms et al. (1975) presented a classification of bed forms in which bed form formation is a function of energy (dependent upon the energy source and water depth), and grain size, where a larger grain size effectively reduces the amount of energy affecting the bed (Table 4). The hierarchy of bed form formation by increasing energy includes ripples, megaripples, and sand waves. Within the ripple classification, a gradation exists from short-crested (0- to 20-cm wavelength), to medium-crested (20- to 40-cm wavelength), to large-crested (40- to 60-cm wavelength) ripples (Reineck and Singh 1986). Within the megaripple classification, a gradation exists from two-dimensional (straight-crested) megaripples, to three-dimensional or lunate (sinuous-crested) megaripples, to flat (plane) beds (Figures 11 and 12).

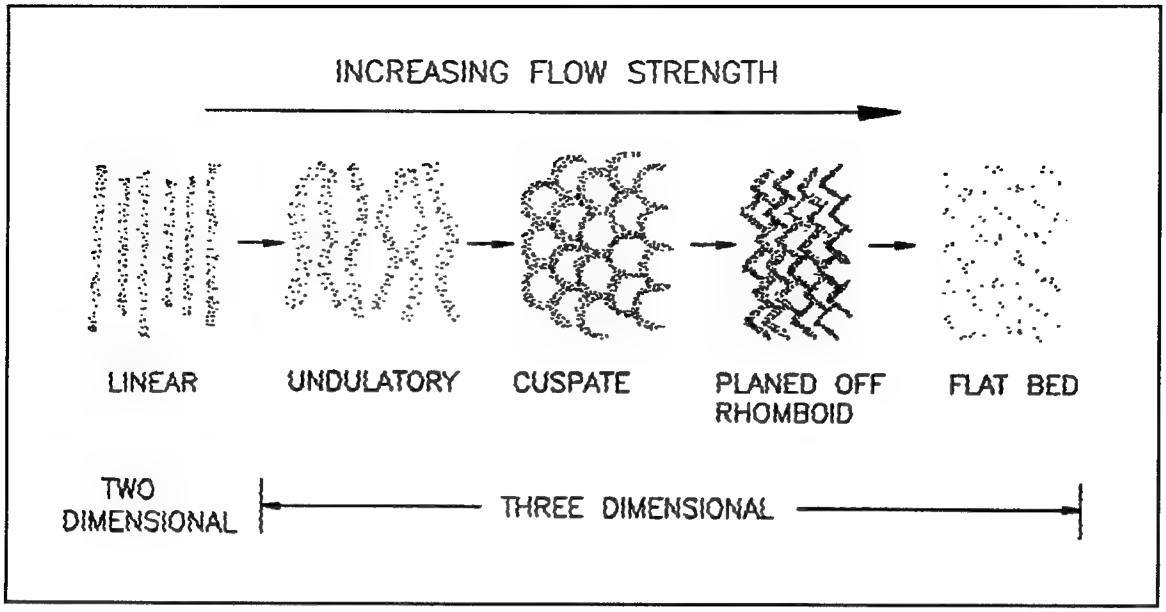

Figure 11. Gradation from two-dimensional to three-dimensional bed forms and flat beds with increasing flow strength (after Reineck and Singh (1986))

Table 4
Hierarchy of Bed Form Formation by Increasing Energy (after Harms et al. (1975))

Parameter	Ripples	Megaripples	Low-energy Sandwaves	High-energy Sandwaves
Spacing	< 60 cm	60 cm - 10 m	> 6 m	> 10 m
Height/Spacing Ratio	Variable	Relatively large	Relatively small	Very small
Geometry	Highly variable	Sinuous to highly three-dimensional	Straight to sinuous	Straight to sinuous
Characteristic Flow Velocity	Low (> 25-30 cm/s, < 40-50 cm/s)	High (> 70-80 cm/s, < 100-150 cm/s)	Moderate (> 30-40 cm/s, < 70-80 cm/s)	High (> 70-80 cm/s, may be 150 cm/s)
Velocity Asymmetry	Negligible to substantial	Negligible to substantial	Usually substantial	Small to substantial

Formation and movement of inner shelf sedimentary features, primarily the smaller scale ripples, are primary methods of inner shelf cross-shore sediment transport. These bed forms are formed only during *turbulent flow conditions* (water flow in which the flow lines are confused and heterogeneously mixed (Bates and Jackson 1984). These turbulent conditions are created by wave and related oscillatory motion, or tide-generated currents near the bottom which roll and creep sediment particles along the sediment-water interface (Reineck and Singh 1986). As sediment particles continue to move from the trough to the crest on both sides, ripples eventually form. As velocity increases and greater amounts of sediment

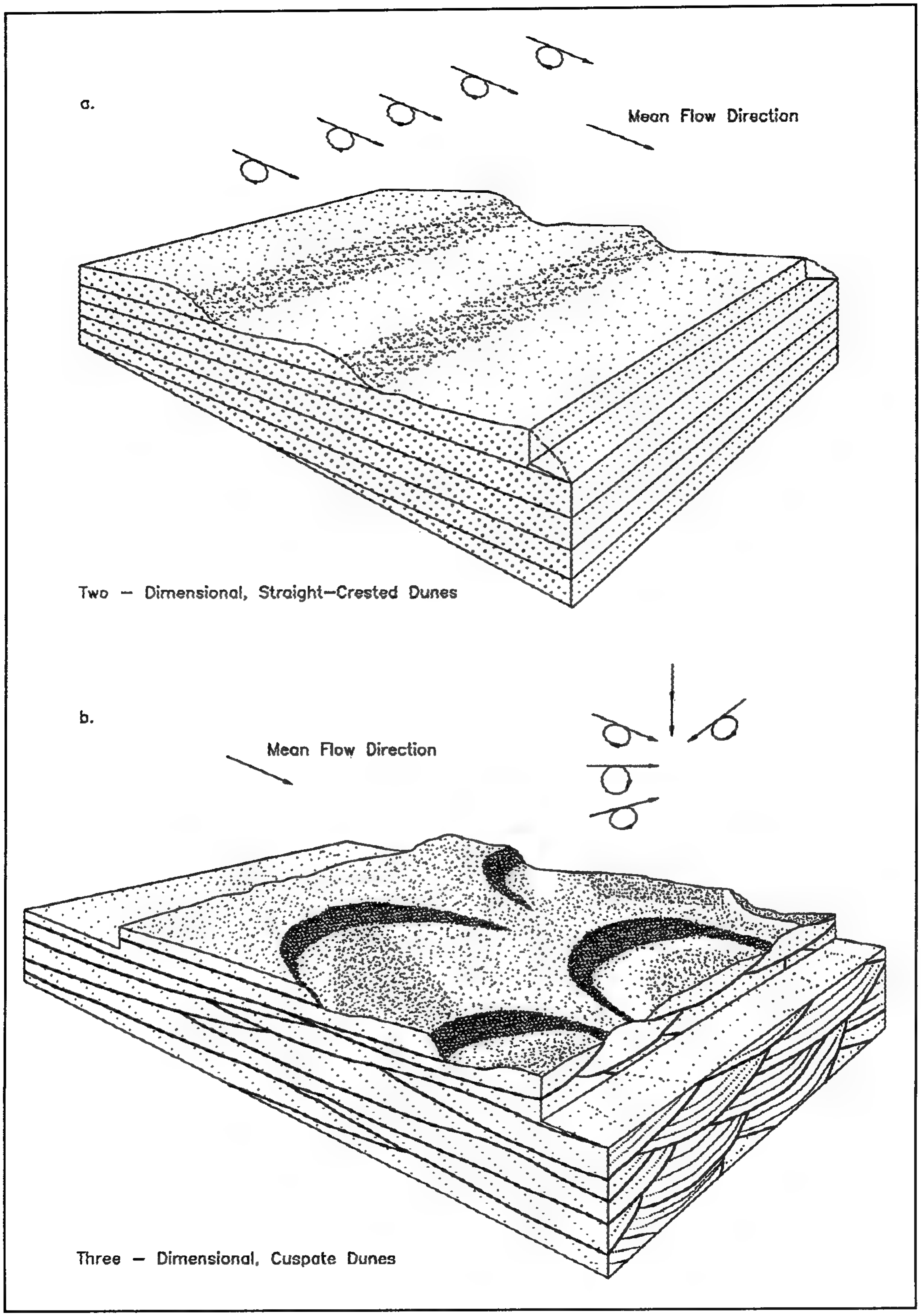

Figure 12. Two-dimensional and three-dimensional bed forms. Vortices and flow patterns are shown by arrows above the dunes (after Reineck and Singh (1986))

are added to the ridge, ripple height continues to increase with velocity until a point where height decreases and length increases.

Bedding theory. *Bedding* is defined as the signature of migration of a surficial *bed form*, or a morphologic feature having various systematic patterns of relief which is created by the conditions of flow at the dynamic interface between a body of cohesionless sediment particles and a fluid (Davis 1983). Many authors have stated that bed form migration produces internal stratigraphic records in subsurface sediments. These records provide clues to the processes, magnitudes, and directions of sediment transport that formed them (Nittrouer and Sternberg 1981, Swift et al. 1983). In other words, a specific process with a given magnitude and direction of energy will produce a unique subsurface stratigraphic record. The reader is referred to Reading (1978), Allen (1982), and Reineck and Singh (1986) for comprehensive discussions of stratigraphic signatures of migrating sedimentary features.

Generally, there exist two classes of bedding; horizontal and cross-bedding. Horizontal bedding is characterized by parallel beds graded at any angle, usually resulting from flat bed sediment migration or the migration of sediment where no bed forms occur.

Cross-bedding, which is the most common type of bedding encountered on the inner shelf, is defined as a single layer, or a single sedimentation unit, consisting of laminae that are inclined in a direction similar to the principal surface of sedimentation. This sedimentation unit is separated from adjacent layers by a surface of erosion, nondeposition, or abrupt changes in character.

Reineck and Singh (1986) indicate that different types of cross-bedding result from the migration of different types and sizes of bed forms. Two types of cross-bedding shown in Figure 13 include:

a. *Planar cross-bedding* - cross-bedding in which bounding surfaces form more or less planar surfaces. These units are tabular or wedge-shaped.

b. *Trough cross-bedding* - cross-bedding in which bounding surfaces are curved surfaces and the unit is trough-shaped.

Clifton (1976) classifies internal sedimentary structures on the inner shelf into the following three classes:

a. *Planar parallel laminae* (where *lamina* (singular) is a type of bedding defined as the thinnest recognizable layer in a sediment differing from other layers (commonly 0.05 to 0.10 mm thick)).

b. *Medium-scale ripple-foreset bedding* (a *foreset* is a type of bedding thicker than lamina produced by the deposition of sediment on the downcurrent face of a bed form (Bates and Jackson 1984).

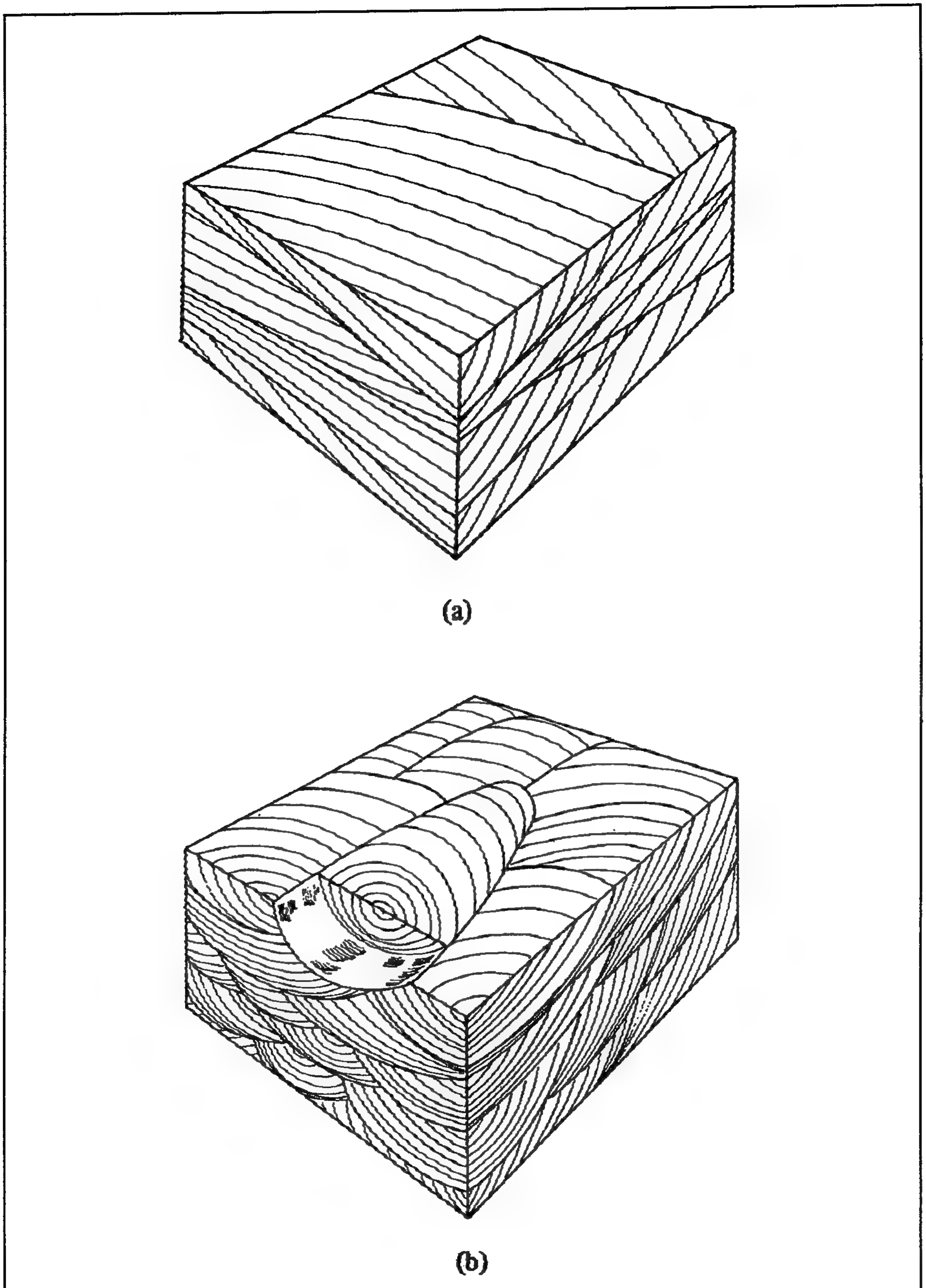

Figure 13. Block diagrams showing planar (a) and (b) trough cross-bedding as seen in horizontal, transverse, and longitudinal sections (after Reineck and Singh (1986))

c. *Small-scale ripple-foreset bedding.*

These three classes of bed forms can form from either wave- or tidal-generated currents depending on the flow characteristics.

Planar parallel laminae develop in shallow marine sands by:

a. *Sheet flow* caused by the consistent flow of sand over a flat bed during high-energy conditions (Davis 1983).

b. Migration of long-crested ripple forms accompanied by a slow rate of sediment accumulation.

Deepwater sheet flow results from the high energy oscillatory flow of large long-period waves (Clifton 1976), and the currents usually associated with geostrophic or downwelling currents. Shallow-water sheet flow results from intense wave activity close to the shoreline and may show evidence of shear sorting of particles of different size, density, or shape (less velocity is needed to form sheet flow in fine sand than in coarse sand). Other sedimentary structures associated with sheet flow include mica laminae, convex-up shells, and little to no bioturbation due to wave reworking.

The second cause of planar parallel laminae is the migration of ripples accompanied with a slow rate of sediment accumulation known as *slowly climbing ripple stratification* (or the internal structure formed in noncohesive material from migration and simultaneous upward growth of long-crested ripples). Climbing ripple stratification can be produced by either currents or waves (Reineck and Singh 1986) of all periods, but only by medium- to long- period waves (8 to 12 sec) in deeper water. The sedimentary signature of the migration of ripple forms accompanied with a slow rate of sediment accumulation includes poorly defined climbing ripple foresets, shell lag deposits, concave up shells due to their tumbling over ripple crests, and bioturbation.

Medium-scale ripple foreset bedding is characterized by 6-cm-thick foreset units in medium to coarse sand, which form due to the migration of cuspate (three-dimensional) megaripples or the migration of long-crested ripples if a rapid sedimentation rate is present. Lunate megaripple migration produces cross-bedding, while long-crested ripple migration produces more *tabular units* (said of the shape of a sedimentary body whose width/thickness ratio is greater than 50 to 1, but less than 1,000 to 1). The foresets of medium-scale foreset bedding are oriented onshore in the direction of wave propagation suggesting the landward transport of sediment associated with orbital asymmetry.

Small-scale ripple foreset bedding is the most common structure near the sediment water interface, but has a low preservation potential. This type of bedding is characterized by foreset units less than 6 cm thick and is produced by the migration of irregular asymmetrical wave ripples (to be described in the following section) or by the migration of small-scale

ripples during rapid sediment accumulation. Bedding planes dip onshore from wave-generated currents, while bedding associated with unidirectional currents dips either onshore or offshore (Clifton 1976).

Ripple symmetry. Inner shelf ripples can be symmetrical or asymmetrical. Symmetrical ripples have similar side slopes and are usually produced by waves and associated bidirectional currents of near similar magnitudes (Reineck and Singh 1986).

Asymmetrical ripples, or ripples with different side slopes, are formed by bidirectional currents of different magnitudes (Reineck and Singh 1986). These bidirectional currents can be formed by both wave and tidal-generated currents. Asymmetrical wave ripples occur especially in the surf zone and shallow water under long period low waves, as the oscillatory flow of water particles tends not to occur in a closed orbit. Net transport of sediment occurs in the direction of wave propagation. Therefore, there is significant unidirectional sediment movement associated with asymmetrical wave ripples. Although both asymmetrical wave ripples and current ripples have unequal side slopes, asymmetrical ripples bifurcate while current ripples do not. Since the formation of bed forms on the inner shelf environment is dominated by wave activity, the following discussion concerns *wave ripples* (ripples formed by wave-generated currents, also known as oscillation ripples) rather than *current ripples* (ripples formed by tidal-generated currents).

Sediment movement in symmetrical wave ripples is a function of wave orbitals at the water surface, which flatten towards the bottom eventually having only horizontal, and not vertical, movement. These ripples are essentially straight-crested, have pointed crests, rounded troughs and frequently show bifurcation. The occasional rounding of crests is a result of the reworking of ripples as the current field changes characteristics. The internal structure of wave symmetrical ripples is characterized by chevrons indicating two directions of transport (*chevron bedding* slopes away from the crest and toward the trough of a ripple at equal angles). A more detailed discussion of internal structure characteristics of wave-ripple bedding can be found in Boersma (1970) and Reineck and Singh (1986).

Clifton (1976), building on the work of Inman (1957) and Dingler (1974), stated that the prediction of symmetrical ripple size, which is gradational, is based on grain size, orbital velocity, and wave period. Three types of symmetrical ripples include (Figure 14):

a. *Orbital ripples*, which form under short-period waves and have ratios between orbital diameter/grain diameter (d_o/D) which are less than 2,000 (where ripple wavelength is dependent upon the length of orbital diameter of the oscillatory current and is independent of grain size).

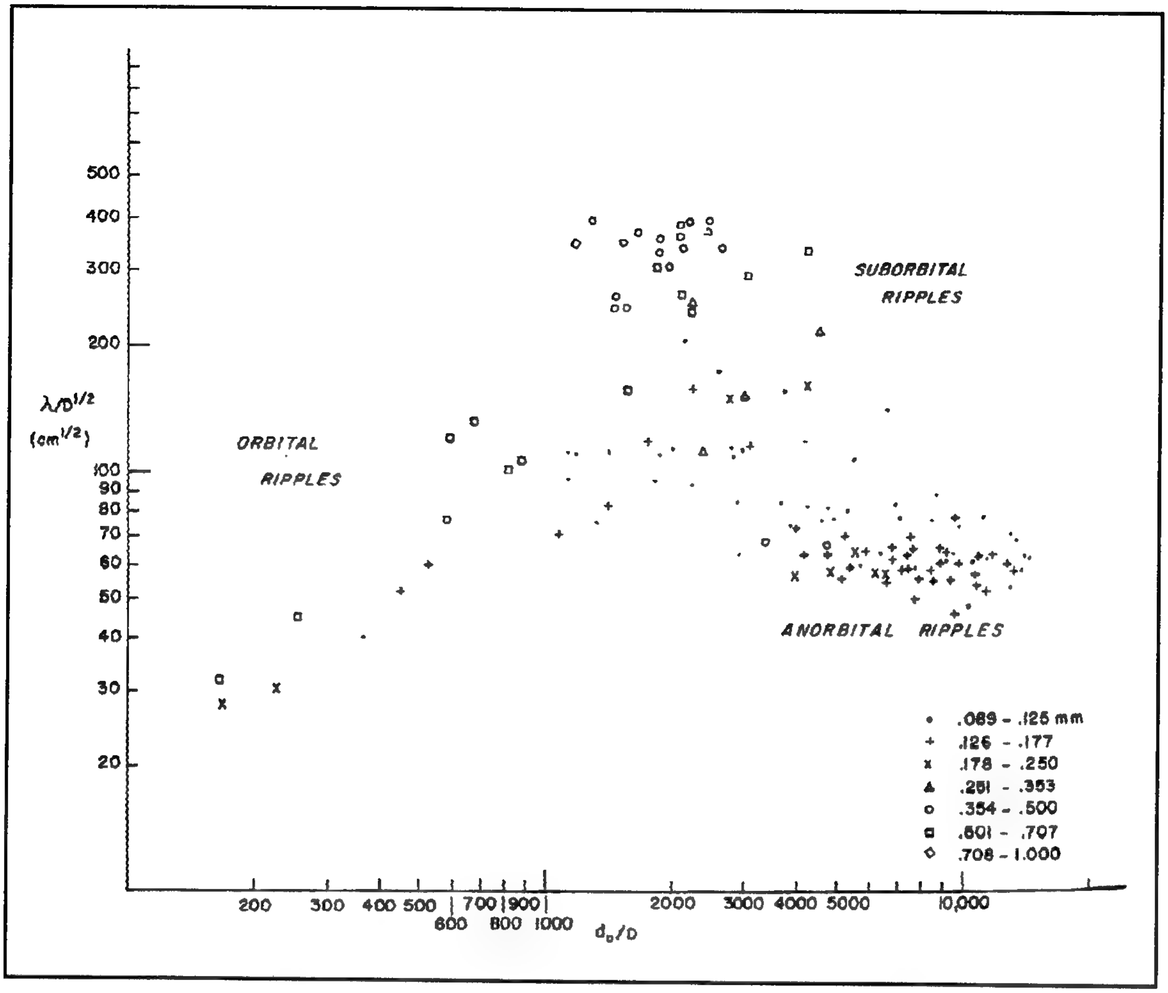

Figure 14. Classification of symmetric and reversing ripples based on the ratio of ripple length to square root of grain diameter ($\lambda/D^{1/2}$) and ratio of orbital diameter to grain diameter (d_0/D) (after Clifton (1976)) based on data from Inman (1957) and Dingler (1974))

b. *Suborbital ripples,* which form under longer period waves and have d_0/D ratios between 2,000 and 5,000 (wavelength increases with larger grain size but decreases with increasing orbital diameter).

c. *Anorbital ripples,* which are associated with waves of very large orbital diameter and have d_0/D ratios greater than 5,000 (wavelength depends on grain size and is independent of orbital diameter).

Reversing ripples, which are considered asymmetrical, have d_0/D ratios between 6,500 and 13,000 (Inman 1957).

In comparing symmetrical and asymmetrical wave ripple size, Clifton (1976) states that symmetrical wave ripples form where maximum bottom orbital velocity is less than 1 cm/sec, while asymmetrical wave ripples form when maximum bottom orbital velocity is greater than 5 cm/sec.

Symmetrical wave ripples, which tend to form in deeper water, do not migrate and thus produce no stratigraphic record. Asymmetrical wave ripples tend to form in shallow water. In addition, symmetrical wave ripples have a poorer preservation potential than asymmetrical ripples, as asymmetrical wave ripples migrate. Komar (1974) indicates that ripple spacing of symmetrical wave ripples increases landward under short-period waves but decreases landward under longer-period waves.

Reineck and Singh (1986) discuss the formation of ripples as a function of water depth and wave period. For wave periods of 2-4 sec, ripples form out to a water depth of -25 m. Symmetrical suborbital ripples are the dominant ripple type for these periods. No asymmetrical ripples form and there exists a limited occurrence of flat beds. For wave periods of 5-8 sec, ripples form out to a water depth of -100 m and are dominated by suborbital symmetrical ripples with some anorbital ripples forming at higher velocities in fine- to medium-grained sand. Flat beds form under large wave conditions except in coarse sand. For wave periods of 10 to 15 sec, ripples form to a water depth of -300 m. In deep water, symmetrical suborbital ripples form in coarse sand while anorbital ripples form in fine sand. It is possible that lunate ripples and flat beds form in medium to coarse sand at higher velocities. Reineck and Singh (1986) also note that maximum velocity, velocity asymmetry, and grain size increase in a landward direction.

Wave-formed sedimentary structures. Clifton (1976) presents a model concerning the origin and interrelationship of wave-formed sedimentary structures. Data collected from southern Oregon (high energy), southeast Spain (relatively low energy) and Willapa Bay, Washington (low energy), and previously collected data from Komar and Miller (1973, 1974), Komar (1974) and Dingler (1974) form the basis for this conceptual model. The processes responsible for these structures include:

a. Wave parameters including height, period, maximum bottom orbital velocity, and change in maximum bottom orbital velocity.

b. Fluid factors (density,viscosity).

c. Flow factors (existing mean currents).

d. Bottom configuration factors (water depth over all and local slope).

e. Sediment factors (grain size diameter, sorting, density, and shape).

f. Oscillatory currents just above the boundary layer.

g. Length of oscillatory water movement.

h. Velocity asymmetry of oscillatory currents.

Arnott and Southard (1990), in a collinear oscillatory and combined flow water tunnel with a wide range of component speeds and an oscillation period of 8.5 sec, have produced stability fields for wave-generated bed forms in very fine sand. Figure 15 shows that different types of bed forms and resulting internal stratigraphy are formed according to different wave oscillatory speeds, which are greater closer to shore and reduce in an offshore direction.

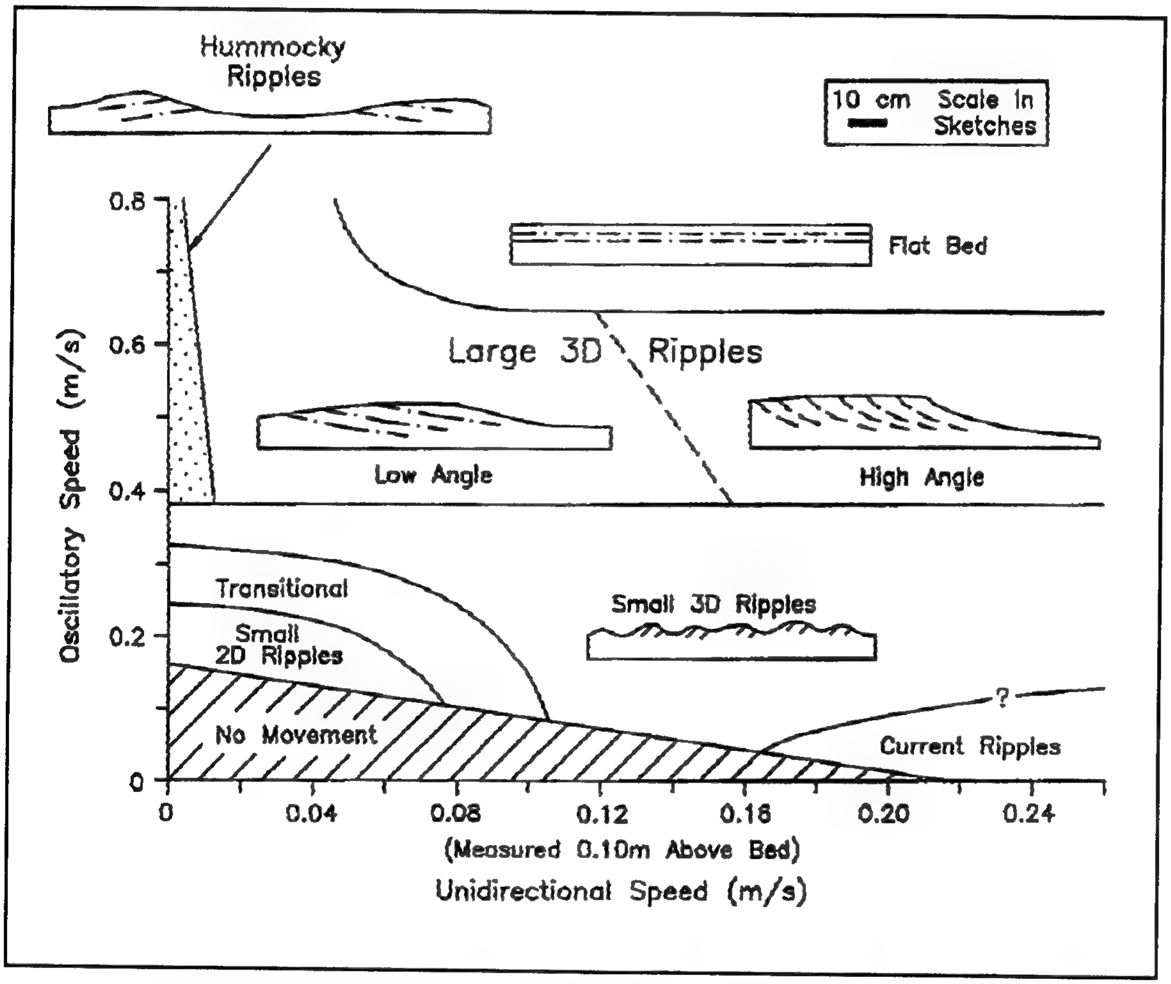

Figure 15. Stability fields for bed forms produced in very fine sand in collinear combined-flow water tunnel. Velocities were measured at 0.10 m above the bed. Note that "2D" considers a two-dimensional (straight-crested, which is usually representative of low energy conditions) bed form, while "3D" considers a three-dimensional (sinuous-crested, usually representative of high energy conditions) bed form (from Arnott and Southard (1990))

Inner Shelf Stratigraphy

Cross-shore stratigraphic sequences

Numerous authors (see Appendix B, "Sedimentary Features and Stratigraphy References") have identified cross-shore sequences of sedimentary structures and resulting stratigraphy. Clifton (1976) documents the following typical sequence of sedimentary structures for the Oregon coast inner shelf resulting from wave-induced oscillatory flow (Figure 16), beginning offshore and moving landward:

a. Inactive zone.

b. Active asymmetric ripples.

c. Long-crested asymmetric ripples.

d. Irregular asymmetric ripples.

e. Asymmetric cross-ripples.

f. Megaripples.

g. Flat bed.

Similar sequences were also found in Australia by Boyd (1981).

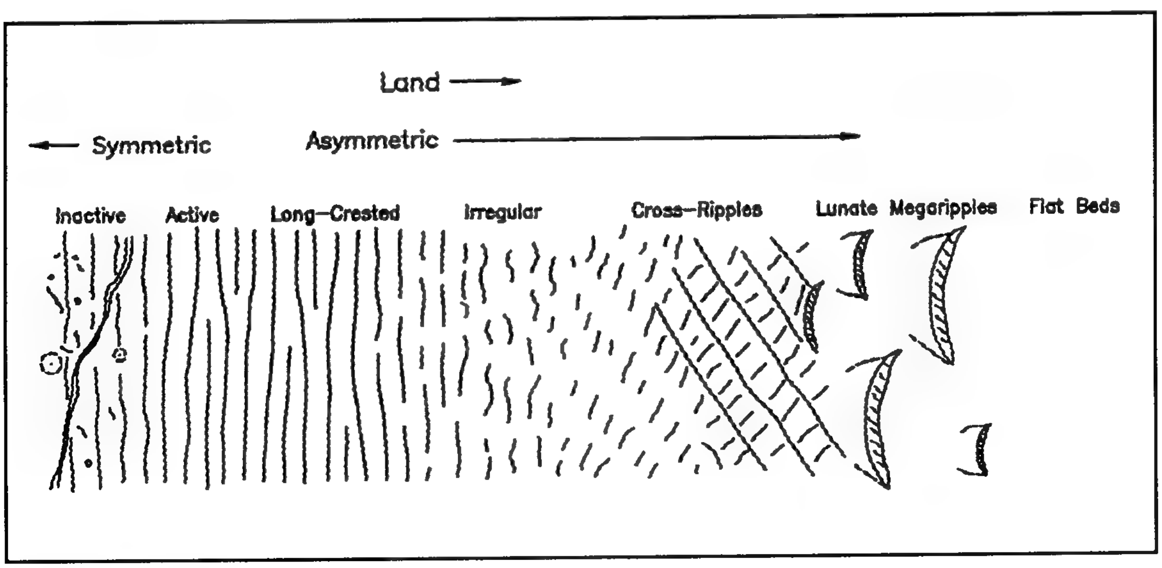

Figure 16. Cross-shore sequence of structures commonly found off the coast of southern Oregon (after Clifton (1976))

Howard and Reineck (1972) defined a cross-shore sequence of internal stratigraphic structures. In addition to a seaward-fining sediment grain size trend, they found that physical sedimentary structures decrease and biogenic structures increase in a seaward direction due to increasing depth and position of the wave base. Howard and Reineck (1981) also examine and describe the primary physical sedimentary structures and compare a high-energy sequence at Port Hueneme, California, with a low energy, tide-dominated sequence at Sapelo Island, Georgia.

Howard and Reineck (1981) describe three facies associated with the Port Hueneme, California, beach-to-offshore depositional stratigraphic sequence. This sequence includes *nearshore*, *transition,* and *offshore* facies. The *nearshore* facies (+3.0-m to -9.0-m water depth)(inclusive of the *foreshore* facies from +3.0 m to 0.0 m, and the *inner shelf* (shoreface) facies from 0.0 to -9.0 m) is composed primarily of parallel and cross-bedded homogeneous sand, and small-scale wave ripple laminae, while bioturbation is only locally significant. Rounded rock-fragment pebbles are present both individually and as layers in the foreshore and more commonly in the swash zone. Alternating layers of coarse and fine sand are locally present. Heavy minerals are abundant throughout and enhance the expression of physical sedimentary structures.

In sections of parallel laminated sand in the nearshore facies, the dip is very low (3 deg) and therefore dip directions cannot be specified from cores. Individual laminae pinch out at erosional contracts suggesting that these are wedged-shaped laminae sets. Thickness of individual parallel sets varies from 1 to 12 mm, with their average thickness being 1-2 mm. Cross-bedded sand is characterized by sets 10 to 30 cm thick with individual laminae up to 2 cm thick. This sedimentary structure is found only in the nearshore facies, and within this facies, increases with decreasing water depth. Cross-bedding is most abundant in the vicinity of the mean low water line and is commonly associated with coarse sand, and alternating sets of coarse and fine sand. Small-scale wave ripple laminae are restricted mainly to the nearshore facies. Ripples are present on the bottom, but were not preserved in cores. Bioturbation was practically nonexistent out to a water depth of -6.3 m as wave activity dominated the sedimentary sequence. Sand dollars were present in water depths from -6.5 to -8.7 m. No shells or shell fragments were found in the nearshore facies (Howard and Reineck 1981).

The *transition* facies (-9.3-m to -18.7-m water depth) is a zone of fine sand and silty sand characterized by an increase in biogenic over physical structures that are commonly preserved as laminated-to-burrowed beds. This laminated- to-burrowed bed sequence is also described by Howard (1972), Howard and Reineck (1972), Golding and Bridges (1973), and Bourgeois (1980). Howard and Reineck (1981) state that wave-ripple bedding and parallel laminae are important structures in this facies. *Hummocky cross-stratification* laminae are defined as laminae which are both concave up (swales) and convex up (hummocks), possess many undulating erosion surfaces, and dip into the swales at angles of approximately

15 deg (to be described in detail in the next section). Hummocky cross-stratification laminae are probably the most persistent physical sedimentary structure in this facies, with small-scale oscillation-ripple laminae second in abundance. Cross-bedding, pebbles, and heavy-mineral laminae are not present. This facies contains stratigraphic structures of both the offshore and nearshore zones. The cross-shore transition between biogenic and physical structures indicates fluctuation of wave energy. The onshore limit of this area is most likely normal wave base, while the offshore limit is storm wave base. No shells or shell fragments were present in this facies (Howard and Reineck 1981).

In the *offshore* facies at Port Hueneme (> -18-m water depth), the primary texture is sandy silt and bioturbation is the dominant sedimentary structure (Howard and Reineck 1981). Energy decreases with increasing water depth, which results in increasing amounts of biogenic activity and a fining of grain size in an offshore direction. Biogenic processes affect up to 90-100 percent of this facies due to the following:

a. Slow rates of sedimentation.

b. Brief storm events.

c. Long periods of relative quiescence.

Remnant parallel laminae are the only physical sedimentary structures present. Shells and shell fragments are abundant. Direct or indirect effects of storms are rare.

In comparing the stratigraphy of the inner shelf off Port Hueneme, California, and Sapelo Island, Georgia, Howard and Reineck (1981) found several differences in the sedimentary sequences resulting from different wave characteristics (as the tidal range for the two areas is similar). A major difference between sedimentary sequences at the two sites was the water depth at which facies boundaries occur. At the Port Hueneme, California, site, the foreshore-inner shelf boundary is distinct as the parallel laminated sand of the foreshore facies is replaced by large-scale cross-bedding, and small-scale ripple laminae of the inner shelf facies. At the Sapelo Island, Georgia site, a distinction between the foreshore/inner shelf boundary could not be made because the parallel laminated sand of the foreshore facies continues as the dominant sedimentary structure well into the upper inner shelf facies.

Thickness of the inner shelf facies was also different between the two sites. At Sapelo Island, the inner shelf is 250 m wide and 2 m thick. The upper inner shelf is characterized by parallel laminated sand, and the lower inner shelf is characterized by small-scale ripple laminae. In contrast, the Port Hueneme inner shelf is 300 m wide and 9 m thick. Large-scale cross- bedding as well as parallel laminated sand and small-scale ripple laminae occur on this inner shelf.

Additional differences between the two sites include the transition zone, which is between the -2.0- and -5.0-m water depths at Sapelo Island, and between the -9.3 and -18.7-m depths at Port Hueneme. Offshore facies are characterized by the presence of *palimpsest sediments*, defined as reworked sediments of the continental shelf, and occur seaward of -5.0 m at the Sapelo Island site, and seaward of -18.7 m at the Port Hueneme site. In addition, storm units (parallel laminated to burrowed beds, separated by erosional contacts) are more clearly developed at the Port Hueneme site sequence.

In a study of Topsail Island, North Carolina, Schwartz, Hobson, and Musialowski (1981) collected data supporting the subdivision of the inner shelf into *upper*, *middle*, and *lower inner shelf* zones. These zones correspond to the inner shelf, transition zone, and offshore facies attributed to the Sapelo Island, Georgia, coast site by Howard and Reineck (1981). Each zone is related to a particular set of nearshore processes and resulting stratigraphical characteristics. The *upper inner shelf* is dominated by surf conditions (including longshore currents) and maximum wave shoaling effects just prior to breaking. The approximate water depth range of the upper inner shelf is estimated to be between 0.0 m and -2.0 m based on sedimentary structures, sediment grain size characteristics, and changes in profile shape). Stratigraphically, the upper inner shelf is characterized by subhorizontal laminae and very low-angle, thinly laminated units, and by local occurrences of inverse textural grading.

The *middle inner shelf* (approximate water depth from -2.0 to -4.0 m), is dominated by relatively strong shoaling effects and coastal currents that produce significant downward scour and sediment transport during storm events. This facies is dominated by subhorizontal laminae, trough cross-bedding, low-angle foreset laminae, and minor bioturbation structures. The *lower inner shelf*, (water depth from -4.0 m to -6.5 m), is slightly to moderately affected by fair-weather waves, is stratigraphically dominated by subhorizontal to low-angle laminar bedding, small-scale trough or ripple bedding, and has moderate to locally abundant bioturbation. Normally, graded beds, although sometimes poorly defined, occur throughout the inner shelf.

Storm-related stratigraphy

Numerous authors have identified storms as controlling sedimentation and stratigraphy of the inner shelf (Appendix B, "Significant (Storm) Event References"). Smith and Hopkins (1972) state that erosion of the continental shelf by severe storms ranges from a few millimeters to centimeters; sediment is transported off the continental shelf into deeper water. Smith and Hopkins (1972) suggest that deposits are layered, and perhaps graded by storms as sands are covered by silt that settles out in suspension after the storms.

Storm-influenced bedding

Types of storm-influenced bedding include the following:

a. *Hummocky cross-stratification* - defined as laminae which are both concave up (swales) and convex up (hummocks), possessing many undulating erosion surfaces, and dip into the swales at angles of approximately 15 deg (Brenchley 1985, 1989). The laminae are oriented 360 deg, indicating that current orientation fluctuates over an entire 360-deg circle. The beds, which thin over hummocks and thicken over swales, appear similar when viewed from two faces perpendicular to one another. Therefore, three-dimensional views are required to correctly identify hummocky cross-stratification (Brenchley 1985, 1989).

b. *Beds of laminated silt,* usually only a few centimeters thick at most, which fine upwards.

c. *Beds similar in nature to turbidites* (where turbidites are defined as a bedding sequence formed by a *turbidity current* or a bottom-flowing current laden with suspended sediment and possessing a density greater than that of the water which moves slowly down a subaqueous slope (Bates and Jackson 1984)). These beds show graded, parallel laminae or ripple drift lamination, commonly formed below the wave base.

Hummocky cross-stratification, also known as *truncated wave ripple laminae* (Campbell 1966, 1971), is of utmost importance in the study of storm deposits on inner shelf sedimentation/stratigraphy patterns. Studies concerned with this subject include Campbell (1966, 1971), Harms (1975), Hamblin and Walker (1979), Bourgeois (1980), Allen (1982), Dott and Bourgeois (1982), Swift et al. (1983), Walker, Duke, and Leckie (1983), Brenchley (1985, 1989), Duke (1985, 1987, 1990), Greenwood and Sherman (1984), Klein and Marsaglia (1987), Nottvedt and Kreisa (1987), Swift and Nummedal (1987), Arnott and Southard (1990), Higgs (1990), Southard and Boguchwal (1990), and Duke, Arnott, and Cheel (1991).

Hummocky cross-stratification requires an increase in seaward sediment transport, and entrainment and deposition of sand on the continental shelf above the wave base by storm-generated currents and waves (Brenchley 1985). This bedding is usually formed by accretion as laminae thicken over crests. However, some hummocky cross-stratification bedding is produced by erosion when sediment is eroded from the hummocks and is deposited and thickens in the swales. Brenchley (1985) questions whether wave oscillatory currents or a combination of wave oscillatory and unidirectional currents are needed to produce hummocky cross-stratification.

Arnott and Southard (1990) state that meter-scale, isotropic hummocky cross-stratification is likely formed by large three-dimensional symmetrical wave ripples produced by purely oscillatory flows and very strongly oscillatory-dominant combined flows of storm waves. They documented that the sedimentary response of the inner shelf from pure oscillatory flow at low speeds was small symmetrical vortex ripples. At higher current velocities large, three-dimensional, round-crested bed forms with heights to 20 cm and spacings of decimeters to meters resulted.

Hummocky cross-stratification varies with distance from shore and water depth (Arnott and Southard 1990). As energy decreases in an offshore direction, hummocky cross-stratification laminae tend to be less deeply incised and dip at a lower angle. At nearshore locations, there is a greater presence of wave ripples, and beds are lenticular (resulting from high energy) and tend to erode at the top. At offshore locations where the energy is less, the beds become tabular. In addition, wavelength and height of hummocks are likely to decrease in an offshore direction.

Arnott and Southard (1990) found that superimposition of a steady current with oscillatory motion produced significant changes in bed state. Even a weak current caused bed forms to become asymmetric and migrate; most of the combined-flow bed forms contained downstream-dipping cross-stratification. Changes in the morphology of the ripples were profound as currents increased. Currents of only 1-5 cm/sec, superimposed on oscillatory flows of 40-60 cm/sec, produced downstream-dipping low-angle hummocky cross-stratification. For currents exceeding 13 cm/sec, hummocky cross-stratification occurred and dip angles were formed near the angle of response (similar in morphology to high-angle hummocky cross-stratification as described by Nottvedt and Kreisa (1987). At higher oscillatory speeds (60-80 cm/sec), any non-negligible current washed the ripples away, replacing them with a flat bed. However, Arnott and Southard (1990) state that a core current exceeding 95-110 cm/sec is needed to form large ripples exhibiting moderately steep internal laminae in very fine sand.

Examples. Greenwood and Hale (1980), in a study at New Brunswick, Canada, using depth of disturbance rods, found that the depth of activity at a bar is proportional to storm intensity. The seaward side of the bar crest, which had maximum values of bed-level change due to large wave heights, asymmetric oscillatory motion, and rip currents, eroded up to 35 cm. Meanwhile, the trough at the foot of the landward slope eroded up to 37 cm due to scour by longshore currents. Accretion of up to 12 cm occurred on the upper part of the landward slope in response to a decrease in wave height due to breaking waves and increased water depth. In addition, accretion of up to 21 cm occurred on the upper seaward slope of the bar, thus steepening both slopes and producing a seaward displacement of the bar crest. Overall, the bar eroded during the storm, and sediment was transported in multiple directions through megaripple migration. However, net transport of sediment was in an offshore direction.

Schwartz, Hobson, and Musialowski (1981) distinguished between fair-weather and storm bedding features. They found that storm sequences are marked by:

a. Beds with sharp lower contacts.

b. Normal textural grading (fining of sediment grain size in an upward direction).

c. Laminae bedding throughout or upward transition from laminated bedding at the base to bioturbation in the upper part of the sequence.

Studies by Curray (1960), Hayes (1967c), and Morton (1981) as reviewed by Nummedal and Snedden (1987) show that fine sand moves offshore from the inner shelf during storms and hurricanes. Nummedal and Snedden (1987) summarize that once transported to the continental shelf, little sediment is returned by post-storm flow. The primary sediment source is the portion of the inner shelf between mlw and the break in slope onto the more gently dipping continental shelf. These sediments are redeposited as thin-graded, centimeter-thick, fining-upward, sand bed sequences with sharp erosional bases on an otherwise muddy shelf. Hummocky cross-stratification is present. Hayes (1967c), who studied inner shelf sedimentation caused by Hurricane Carla (September 1961) documented that these beds have a sharp upper contact, suggesting that some erosion occurred after the Hurricane Carla deposition. The beds have a scoured sole-marked base and are floored by a coarse lag of pebbles or shell fragments. Hummocky cross-stratification is common. This suggests that little sand is returned onto the inner shelf and beach from the inner shelf after a hurricane.

In measuring bed level changes during a storm, Green et al. (1988) noted that bed changes at the -8-m depth included 6 cm of accretion over 4.5 days of low-energy flow associated with currents as measured with a digital sonar altimeter prior to the onset of the storm. During the initial phase of the storm, 5 cm of scour was followed by 15 cm of rapid accretion. This accretion was coincident with the organization of surface waves into long-period swell, and maximum accretion was coincident with the most highly skewed waves. Onshore sediment transport correlated strongly with erosion of the bed, and offshore transport with accretion of the bed.

Gagan, Chivas, and Herczog (1990) showed that Cyclone Winifred (1 February 1986) produced a normally graded, mixed terrigenous-carbonate bed sequence 11 cm thick in water depths up to -43 m extending 30 km offshore. Cross-shelf distribution of organic carbon in the sediment indicated that suspended sediment transport was extensive and that the storm layer was the result of the following three sources:

a. Landward transport of reworked, resuspended mid-shelf sediment.

b. Resuspension and settling of inner shelf sediment.

c. Seaward transport of terrigenous sediment in freshwater plumes.

By taking 15-cm cores, Gagan, Chivas, and Herczog (1990) show that on a shelf-wide scale, in the -20- to -40-m water depth, sediment was eroded to a depth of 6.9 cm, and in water depth less than -20 m, sediment was eroded to a depth of 5.1 cm. Particles finer than medium sand were eroded and transported out of the mid shelf.

Gagan, Chivas, and Herczog (1990) found that at least 10-30 percent of inner shelf storm sediment is composed of mid-shelf mud, thus indicating the landward movement of fine material. In summary, Gagan, Chivas, and Herczog (1990) support other findings that significant storms are capable of sporadic but efficient cross-shelf transport of suspended sediment.

Wright et al. (1991) and others (Swift et al. 1983; Niedoroda, Swift, and Hopkins 1985; Niedoroda, Swift, and Thorne 1989) concur with Gagan, Chivas, and Herczog (1990) that the inner shelf is dominated by storm flows, which produce a fining sequence of grain size in an offshore direction, and storm beds including hummocky cross-stratification and storm-graded bedding.

Wright et al. (1991), using a digital sonar altimeter, also documented bed-level changes of 15 cm at 8 m due to a 'Northeaster' storm. This increase is inferred to be a result of offshore migration of sediment lobes possessing abrupt leading edges, which migrate well seaward of the -8-m depth contour. These lobes are indicative of energetic cross-shelf advection, as opposed to gradual diffusion.

Wright et al. (1991) documented the response of the bed primarily as a result of hydraulic roughness during different weather conditions. Bed response during fair-weather conditions was characterized by pronounced wave-induced ripples, low sediment mobility, and high apparent hydraulic roughness heights (up to 1 cm). During post-hurricane fair-weather conditions, the bed was mantled with redeposited fine sediment and exhibited subtle ripples surmounting irregular ridges and depressions. This morphology yielded the lowest hydraulic roughness of all four cases.

During storm-dominated conditions (wave heights and periods of 3-6 m and 10-20 sec, respectively, and near-bottom wind-driven mean currents of 0.5 m/s) while there were no ripples, a highly mobile plane bed was present. However, strong wave agitation and a thick wave boundary layer resulted in an effective hydraulic roughness moderately larger than that of the ripple-dominated normal fair-weather case. Skin friction and total bed stresses during the storm exceed those of fair-weather conditions by more than an order of magnitude.

Swell-dominated conditions created the greatest hydraulic roughness of all four cases. This was due to the existence of a thick wave boundary layer with subtle ripples on a partially armored bed.

In studies of the ancient geologic rock record, Brenchley (1989) and Duke, Arnott, and Cheel (1991) state that hummocky cross-stratification is part of a storm bed sequence characterized by an eroded base with a gradational top, which includes the following activities (from bottom to top of the sequence) (Figure 17):

a. Waves interact with a relatively weak coast-oblique bottom current to erode the muddy substrate. Simultaneously, shells and shell hash carve tool marks in the mud and are deposited in swales).

b. Coastal sand, moving as bed and suspended load under combined wave and current bottom flow, is eventually transported offshore resulting in the formation of horizontal lamination to low-angle dipping sand (this also results in basal erosion).

c. Formation of hummocky cross-stratification due to reworking of the bed by storm processes.

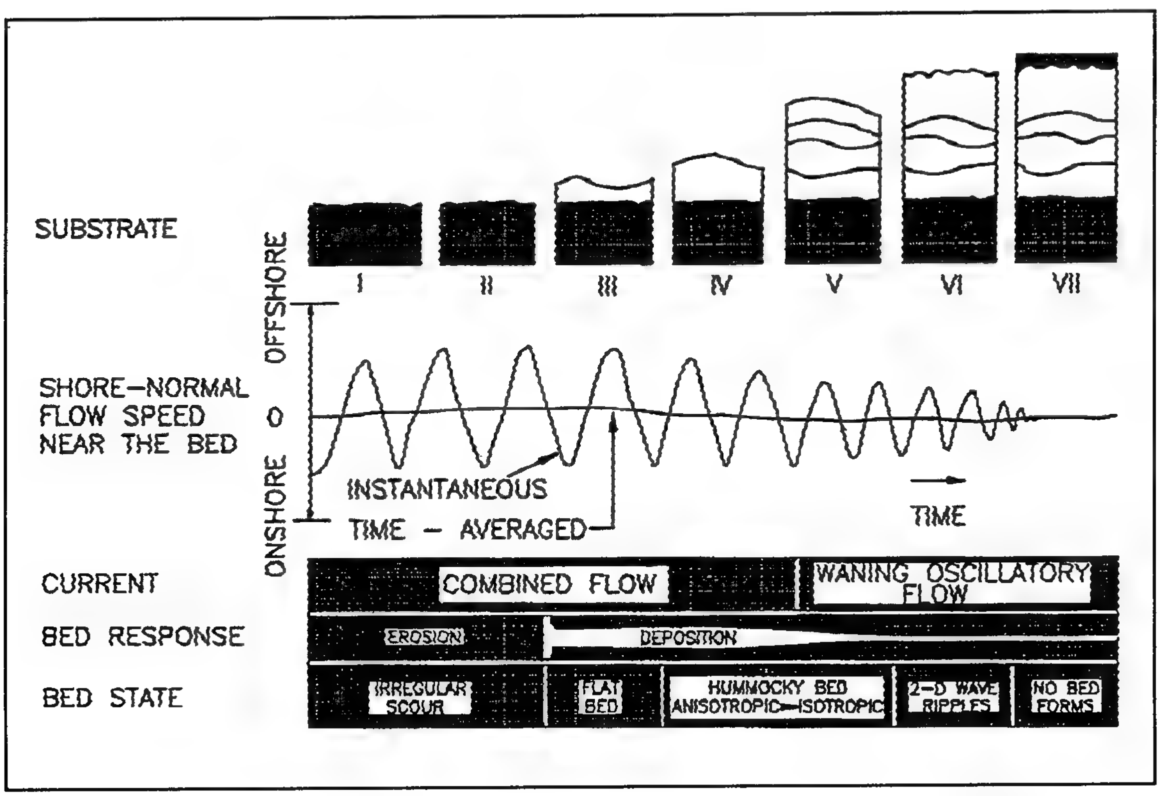

Figure 17. Probable sequence of events producing hummocky cross-stratification on the inner shelf (after Duke, Arnott, and Cheel (1991))

d. As storm processes wane, sand and mud accumulate and are deposited as parallel laminae on top as formed under oscillatory-dominant combined flow (much of it draping over low-relief scours), while megaripples which slowly form and migrate on the still-aggrading substrate may initially produce anisotropic hummocky cross-stratification (bedding properties are different in all directions). Much of the sand is reworked by waves as the bottom current subsides, thus resulting in strongly oscillatory-dominant combined flow and the formation of isotropic (properties are similar in all directions) hummocky cross-stratification. As storm wave motions decrease in speed, a reworked mantle of draping lamination and vortex ripples is formed. Later, the sand is buried by mud and often bioturbated (from Duke, Arnott, and Cheel (1991)).

5 Summary

Nummedal and Snedden (1987) state that during storms and post-storm recovery, large quantities of sand move in cross-shore directions. Large quantities of this sediment may be lost from the beach and from the active profile, thus necessitating beach fill. Much is known about nearshore sediment movement under shoaling waves (Komar 1976); precise documentation of cyclic patterns of surf-zone change (Wright et al. 1979, Nummedal and Snedden 1987), and the well-studied effects of rip currents (Cook and Gorsline 1972, Wright and Short 1984).

However, despite undergoing intense study by geologists and engineers for over a century, there are still many fundamental, unanswered questions about patterns, mechanisms, and rates of beach-shelf sediment interchange. An extensive amount of field work concerning contrasting inner shelf environments is needed (particularly data from cross-shore arrays which provide simultaneous measurements at different depths of near-bottom flows, sediment fluxes, and bed responses). Wright (1987) believes that in determining cross-shore inner shelf sediment transport processes, attention should be placed on field studies and modeling the naturally occurring inner shelf environments. Wright (1987) believes that no one model (or concept) effectively describes inner shelf transport.

Nummedal and Snedden (1987), Wright et al. (1991), and Pilkey (1993) contend that existing models of equilibrium profile development and cross-shore sediment transport are seriously inadequate.

Pilkey et al. (1993) contend that present-day assumptions of the profile of equilibrium concept indicate the following:

a. Sediment movement on the inner shelf is an exceedingly complex phenomenon driven by a wide range of wave, tidal, and gravity currents.

b. The depth of closure does not exist, as evidence shows that large volumes of sand may frequently be moved beyond the depth of closure. These large volumes of sediment moved are often spread over such a large area that standard profiling methods cannot detect this movement.

c. The inner shelf is often not sand rich and in some areas is strongly influenced by the geological framework.

d. The profile of equilibrium equation provides an average inner shelf profile cross section, but does not accurately predict equilibrium profiles at specific inner shelves.

Present-day models concerning inner shelf cross-shore sediment transport and based on the profile of equilibrium equation (Pilkey et al. 1993) do not adequately describe nearshore sediment transport as they say inner shelves can be described and differentiated solely on the basis of sediment grain size and a broadly defined wave climate. However, these models do represent the most up-to-date estimation of inner shelf cross-shore sediment transport and are particularly useful in that they allow an engineer or scientist to explore storm impact on a location using a general approximation of the profile.

Many problems must be understood before we can gain a reasonable understanding of inner shelf and nearshore equilibria/disequilibria and the associated rates of and directions of cross-shore sediment transport (Wright et al. 1991). A goal for the coastal engineering community should be "to devise a more universal conceptual framework capable of better accounting for inner shelf transport, erosion, and deposition in time and space" (Wright 1987). Accomplishing this goal would help to do the following:

a. Garner a better understanding of the physical oceanography of the inner shelf, including the vertical segregation of flows and cross-shelf variations of these flows.

b. On a morphodynamic perspective, study the bottom boundary layer processes that provide the connecting link between hydrodynamics and resulting morphologic change via sediment transport.

c. Study the environmental end members (i.e. other sites) in order to create a comprehensive inner shelf morphodynamic model.

d. Acquire more detailed time series data on near-bottom flow structure, sediment fluxes, bedform behavior, and substrate microstratigraphy. As their empirical base is expanded, so, too, theory and models should be expanded.

e. More accurately predict ripple geometries and their applicability to mixed sediment size distributions and combined waves and currents.

f. Create more realistic paradigms for shelf-nearshore equilibrium that take explicit account of the natural suite of near-bottom flows and of the fundamental roles played by time-varying bed micromorphology.

g. Caution users of any inner shelf models that they must be aware of the limitations of the models and of special conditions that may exist at their project sites.

h. Commence an extensive field measurement and modelling effort not currently underway in North America (Wright 1987, Pilkey et al. 1993).

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Appendix A
Glossary

Bedding - the signature of a migration of a surficial *bed form.*

Bed form - a morphologic feature having various systematic patterns of relief and created by the conditions of flow at the dynamic interface between a body of cohesionless sediment particles and a fluid.

Climbing ripple stratification - The internal structure formed in noncohesive material from migration and simultaneous upward growth of long-crested ripples.

Continental shelf - The gently sloping submerged edge of a continent, extending from the surf zone seaward to a depth of about 130 m, or the edge of the continental slope. The continental shelf is composed of two distinct zones, the inner and outer continental shelf. The shelf is characterized by an average slope of 0.1 deg.

Continental shelf break - The seaward edge of the continental shelf where the bottom begins to descend at a greater angle as part of the continental slope. Average depth of the shelf break is 130 m.

Continental slope - The submerged edge of a continent extending seaward of the continental shelf which is characterized by slopes of 3-6 deg.

Cross-bedding - A single layer, or a single sedimentation unit, consisting of laminae that are inclined in a direction similar to the principal surface of sedimentation. This sedimentation unit is separated from adjacent layers by a surface of erosion, nondeposition, or abrupt changes in character.

Depth of closure - The point on the equilibrium profile beyond which there is no significant net offshore transport of sand even during storm conditions.

Equilibrium profile - The long-term profile which the ocean bed is assumed to conform to based on a particular wave climate and sediment characteristics.

Foreset - A type of bedding thicker than lamina produced by the deposition of sediment on the downcurrent face of a bed form.

Holocene - The Epoch from approximately 10,000 years before present (ybp) to the present, which follows the continental glaciations of the Pleistocene Epoch.

Horizontal bedding - Bedding characterized by parallel beds graded at any angle, usually resulting from flat bed sediment migration or the migration of sediment where no bed forms occur.

Hummocky cross stratification - Laminae which are both concave up (swales) and convex up (hummocks) possessing many undulating erosion surfaces, and dip into the swales at angles of approximately 15 deg.

Inner shelf (inner continental shelf) - The inner part of the continental shelf, also known as the shoreface, extending from the seaward edge of the surf zone to the landward edge of outer continental shelf. This zone is characterized by a normal, strong agitation of the seafloor bed by waves. Slopes of this zone are on the order of 1:200.

Lamina (pl. laminae) - The thinnest recognizable layer in a sediment or sedimentary rock differing from other layers in color, composition, or particle size. Commonly 0.05 to 1.00 mm thick.

Outer continental shelf - The outer continental shelf, the landward limit marking the depth of closure, is only periodically agitated by waves. Slopes of this zone are on the order of 1:2,000.

Palimpsest sediments - Reworked sediments of the continental shelf.

Planar cross-bedding - Cross-bedding in which bounding surfaces form more or less planar surfaces. These units are tabular or wedge-shaped.

Pleistocene Epoch - The Epoch characterized by continental glaciations at North America from approximately 2 million years to 10,000 ybp.

Profile envelope (active) - The range of vertical migration of the profile due to coastal processes including waves and currents.

Quaternary Period - The Period from approximately 2 million ybp to the recent (present) inclusive of the Pleistocene and Holocene Epochs).

Sheet flow - the consistent flow of sand over a flat bed during high energy conditions.

Shoreface - See *inner shelf*

Surf zone - The region characterized by normal and strong agitation of the seafloor bed by the borelike translation of waves following wave breaking.

Trough cross-bedding - Cross-bedding in which bounding surfaces are curved surfaces and the unit is trough-shaped.

Turbulent flow conditions - water flow in which the flow lines are confused and heterogeneously mixed.

Wisconsinan Stage - The most recent and farthest south continental glaciation advancement from approximately 21,500 ybp to 10,000 ybp during the Pleistocene Epoch.

Appendix B
Bibliography with Respect to Topic

This appendix is divided into 12 individual reference lists, each of which concerns a separate piece of evidence of cross-shore sediment transport on the inner shelf.

Individual topics demonstrating evidence of cross-shore sediment transport on the inner shelf include Original Inner Shelf Studies (page B1), Sedimentary Features and Stratigraphy (page B2), Significant (Storm) Events (page B11), Sediment Transport (page B14), Shelf Coastal Processes (page B25), Equilibrium Profile and Profile Adjustment (page B31), Depth of Closure (page B35), Field Research Facility (page B36), Geological Framework (page B39), Comprehensive Studies (page B42), Organic Burrowing (page B43), and Cross-Shore Sediment Transport Model Reference Lists (page B44).

Original Inner Shelf Study References

Purpose

A reference list of some of the original studies concerning cross-shore sediment transport on the inner shelf follows (subject matter of studies is also noted):

a. *Laboratory Studies*
Beach Erosion Board (1947) - Laboratory study of equilibrium beach profiles
Inman and Bowen (1963) - Sediment transport by waves and currents
Rector (1954) - Equilibrium beach profiles

b. *Processes/Hydrodynamics*
Arlman, Santema and Svasek (1958) - Movement of bottom sediment by currents and waves (with radiometric tracer)

Bumpus (1965) - Residual drift along the northwestern United States continental shelf bottom waters
Einstein and Li (1958) - Viscous sublayer along a smooth boundary
Longuet-Higgins and Stewart (1964) - Radiation stress
Manohar (1955) - Mechanisms of bottom sediment movement due to wave action
Shepard and Inman (1950) - Nearshore water circulation related to bottom topography and wave refraction

c. *Equilibrium Beach Profiles*
Bascomb (1951) - Relationship between sand size and beach face slope
Beach Erosion Board (1947) - Laboratory study of equilibrium beach profiles
Bruun (1953) - Forms of equilibrium coasts with a littoral drift
Dietz (1963) - Wave base, marine equilibrium, and wave built terraces
Eagleson, Glenne, and Dracup (1961) - Equilibrium profiles offshore
Fenneman (1902) - Development of the profile of equilibrium
Johnson (1959) - Supply and loss of sand to the coast
Keulegan and Krumbein (1949) - Bottom slope configuration in shallow water and relation to geologic processes
Rector (1954) - Equilibrium beach profiles
Tanner (1958) - The equilibrium beach

d. *Sediment Transport*
Bruun (1962) - Sea level rise as a cause of storm erosion
Caldwell (1956) - Wave action and sand wave migration off the California coast
Cartwright and Stride (1958) - Sand waves on the near shelf
Hall and Heron (1950) - Test of nourishment of the shore by offshore deposition of sand
Inman (1953) - Areal and seasonal variation in beach and nearshore sands in southern California
Inman and Risnak (1956) - Changes in sand level on beach and shelf in southern California
Inman (1957) - Wave-generated ripples in nearshore sands
Shepard (1950) - Beach cycles in southern California
Shepard and Inman (1951) - Sand movement on the southern California shelf
Vernon (1965) - Shelf sediment transport system

e. *Sediments*
Gorsline (1963) - Bottom sediments of the Atlantic shelf and slope of the southern United States
Hayes (1967) - Relation between sediment type and coastal climate on the inner shelf
Shepard (1932) - Sediments of the continental shelves

Uchupi (1963) - Sediments on the continental shelf off the eastern U.S. coast

f. General (Comprehensive Texts)

Johnson (1919). Shore processes and shoreline development

Sverdrup, Johnson, and Fleming (1942). The oceans, their physics, chemistry, and general biology

Sedimentary Features and Stratigraphy References

Purpose

A reference list addressing sedimentation patterns and resulting stratigraphic record of the inner shelf. This reference list also concerns stratigraphic relationships preserved in the ancient rock record.

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Significant (Storm) Event References

Purpose

A reference list addressing the effect of significant events (storms) on cross-shore sediment transport and shelf sedimentation. (References contained in this list are also included in other reference lists included in this bibliography.)

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Purpose

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Shelf Coastal Processes References

Purpose

A reference list addressing coastal processes and associated currents which affect cross-shore transport of sediment on the inner shelf.

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Purpose

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Purpose

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Purpose

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Appendix C
Bibliography with Respect to Topic and Location

General References

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Clifton, Hunter, and Phillips (1971) - Oregon
Dingler (1974) - California
Dingler and Inman (1977) - California
Drake, Kolpack, and Fischer (1972) - California
Drake, Cacchione, and Karl (1985) - California
Greenwood and Mittler (1984) - Canada
Gross, Morse, and Barnes (1969) - Washington, Oregon
Halpern (1976) - Oregon
Howard and Reineck (1981) - California
Hunter, Clifton and Phillips (1979) - Oregon
Inman (1953) - California
Inman (1957) - California
Inman and Risnak (1956) - California (La Jolla)
Inman, Swift, and Duane (1973) - Washington
Kachel (1980) - Washington
Komar, Neudeck, and Kulm (1972) - Oregon
Komar and Miller (1975) - Oregon
Korgen, Bodvarsson, and Kulm (1970) - Oregon

Larsen (1982) - Washington
Miller and Komar (1980) - Oregon
Nittrouer and Sternberg (1981) - Washington
Pilkey et al. (1972) - Oregon
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Shepard (1950) - California
Shepard and Inman (1951) - California (La Jolla)
Smith and Hopkins (1972) - Washington, Oregon
Sternberg (1972) - Washington
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Sternberg and Larsen (1976) - Washington
U.S. Department of Commerce (1984) - California
Vernon (1965) - California

North American Atlantic

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Birkemeier (1985a) - North Carolina (Duck)
Birkemeier (1985b) - North Carolina (Duck)
Birkemeier et al. (1989) - North Carolina (Duck)
Birkemeier et al. (1991) - South Carolina
Bowen (1980) - Canada
Brown, Ehrlich, and Colquhoun (1980) - Southeast Atlantic Coast
Bumpus (1965)
Butman and Folger (1979) - Mid-Atlantic Coast
Butman, Noble, and Folger (1977) - Mid-Atlantic Coast
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Duane et al. (1972)
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Greenwood and Osborne (1991) - Canada (New Brunswick)
Hall and Herron (1950) - New Jersey
Hayden et al. (1975)
Hine and Riggs (1986) - North Carolina
Howard and Reineck (1972) - Georgia
Howard and Reineck (1981) - Georgia

Howd and Birkemeier (1987)
Kraus, Gingerich, and Rosati (1989) - North Carolina (Duck)
Lavelle et al. (1978) - New York
Leffler et al. (1992)
Liu and Zarillo (1987) - New York
Ludwick (1977) - Virginia
Luternauer and Pilkey (1967) - North Carolina
Madsen et al. (1993) - North Carolina (Duck)
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McClennen (1973) - New Jersey
Meisburger and Judge (1989) - North Carolina (Duck)
Niedoroda and Swift (1981) - New York
Osborne and Greenwood (1992) - Canada (Nova Scotia)
Pearson and Riggs (1981) - North Carolina
Pilkey (1968) - Southeast Atlantic Coast
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Riggs (1979) - North Carolina
Riggs (1991) - North Carolina
Riggs and O'Connor (1974) - North Carolina
Riggs et al.. (1986) - North Carolina
Sallenger, Holman, and Birkemeier (1985)
Schmittle (1982)
Schwartz, Hobson and Musialowski (1981) - North Carolina (Topsail Island)
Schwing, Kjerfve, and Sneed (1983) - South Carolina
Shipp (1984) - New York
Seymour (1983) - Virginia (Virginia Beach)
Snyder, Hine, and Riggs (1982) - North Carolina
Snyder and Riggs (1989) - North Carolina
Snyder et al. (1991) - North Carolina
Snyder et al. (1993) - North Carolina
Snyder, Hoffman, and Riggs (in press) - North Carolina
Stauble (1992) - North Carolina (Duck)
Stauble, Garcia, and Kraus (1993) - Maryland
Stefansson, Atkinson, and Bumpus (1971) - North Carolina
Stubblefield, Permenter, and Swift (1977) - New York
Swift and Freeland (1978) - Mid-Atlantic Coast
Swift, Freeland, and Young (1979) - Mid-Atlantic Coast
Swift, Thorne, and Oertel (1986) - Mid-Atlantic Coast
Swift et al. (1981) - New York, Maryland, Massachusetts (Nantucket)
Swift, Thorne, and Oertel (1986)
Swift, Han, and Vincent (1986) - Mid-Atlantic Coast
Twichell (1983) - Massachusetts (Georges Bank)
Vaughn et al. (1987) - North Carolina
Vincent (1986) - Mid-Atlantic Coast
Vincent, Swift, and Hillard (1981) - New York
Vincent, Young, and Swift (1982) - New York
Vincent, Young, and Swift (1983) - New York

Windom and Gross (1989) - Southeast Atlantic Coast
Uchupix (1963) - Mid-Atlantic Coast
Wright et al. (1986)
Wright et al. (1991) - North Carolina (Duck)
Wright (1993) - North Carolina (Duck)

United States Gulf of Mexico

Bernard, LeBlanc, and Major (1962) - Texas
Brooks (1983) - Texas
Dean (1977) - Eastern Gulf Coast
Dupre (1985) - Texas
Forristall, Hamilton, and Cardone (1977)
Gorsline (1963)
Hayes (1967a) - Texas
Hayes (1967b) - Texas
Hayes (1967c) - Texas
Hayden et al.. (1975)
Hill and Hunter (1976) - Texas
Keen and Slingerland (1993a) - Texas
Keen and Slingerland (1993b) - Texas
Morton (1981) - Texas, Louisiana
Morton (1988)
Morton and Winker (1979) - Texas
Murray (1970) - Mississippi
Nummedal and Snedden (1987) - Texas
Smith (1977)
Snedden, Nummedal, and Amos (1988) - Texas

North Sea

Aagaard (1988)
Arlman, Santema, and Svasek (1958)
Bruun (1954)
Morton (1981)
Reineck and Singh (1971)
Swift et al. (1981)
Winkelmolen and Veenstra (1980)

North American Great Lakes

Engstrom (1974) - Lake Superior
Hands (1979)
Hands (1980)
Hands (1981)
Hands (1983)
Hands (1984)

Greenwood and Sherman (1984) - Lake Huron
Greenwood and Osborne (1991) - Georgian Bay
Osborne and Greenwood (1992) - Lake Huron
Stockberger and Woods (1990)

Other Locations

Beydoun (1976) - Eastern Mediterranean Sea
Boon and Green (1989) - Caribbean
Boyd (1981) - Southeast Australia
Channon and Hamilton (1976) - Southwest England
Clifton (1976) - Southeast Spain
Cowell et al. (1983) - Southeast Australia
Field et al. (1981) - Bering Sea
Field and Roy (1984) - Southeast Australia
Figueiredo, Sanders, and Swift (1982) - Brazil
Flemming (1980) - South Africa
Gagan, Chivas, and Herczag (1990) - Southeast Australia
Gao and Collins (1992) - China
Hino, Yamashita, and Yoneyama (1981) - Japan
Hunter, Thor, and Swisher (1982) - Bering Sea
Jago and Borusseau (1981) - France
Kuo, Su, and Liu (1980) - Japan
Kuo et al. (1987) - Japan
Pae and Iwagaki (1985) - Japan
Roy and Stephens (1980) - Southeast Australia
Short (1984)
Wells and James (1981) - South America

Sediment Transport Mechanisms References

These references concern the inner-shelf mechanisms (processes) which result in cross-shore sediment transport. Complete citations can be found in the Coastal Processes reference list in Appendix B.

North American Pacific

Cacchione (1987) - California
Gross, Morse, and Barnes (1969) - Washington, Oregon
Halpern (1976) - Oregon
Korgen, Bodvarsson, and Kulm (1970) - Oregon
Seymour (1986) - California (Torrey Pines, Scripps)

North American Atlantic

Bumpus (1965) - Atlantic Coast
Butman and Folger (1979) - Mid-Atlantic Coast
Beardsley, Butman (1974) - New England
Birkemeier et al. (1989) - North Carolina (Duck)
Crowson et al. (1988 - North Carolina (Duck)
Davidson-Arnott and McDonald (1989) - Canada
Harris, R.L. (1954) - New Jersey
Lavelle et al. (1978) - New York
Madsen et al. (1993) - North Carolina (Duck)
Mason et al. (1984) - North Carolina (Duck)
Niedoroda and Swift (1981) - New York
Osborne and Greenwood (1992a) - Canada (Nova Scotia)
Schwing, Kjerfve, and Sneed (1983) - South Carolina
Seymour (1986) - Virginia (Virginia Beach)
Stefansson, Atkinson, and Bumpus (1971) - North Carolina
Swift, Han, and Vincent (1986) - Mid-Atlantic Coast
Vincent (1986) - Mid-Atlantic Coast
Williams (1976) - New York
Windom and Gross - Southern Atlantic Coast

United States Gulf of Mexico

Forristall, Hamilton, and Cardone (1977)
Hands (1983)
Hands (1991) - Alabama
Murray, 1970 - Mississippi
Smith (1977)
Snedden, Nummedal, and Amos (1988) - Texas
Williams and Meisburger (1987) -New York

North American Great Lakes

Greenwood and Sherman (1984) - Lake Huron
Osborne and Greenwood (1992b) - Lake Huron

Other Locations

Wells and James (1981) - South America

Cross-Shore Sediment Transport References

A reference list documenting references which give evidence of cross-shore sediment transport on the inner shelf (shelf-beach sediment exchange) is divided by regional area as follows: (The reference list entitled "Sediment Transport" in Appendix B addresses additional inner shelf sediment references.)

North American Pacific

Cacchione et al. (1987) - North Carolina
Caldwell (1956) - California (Anaheim)
Drake, Kolpack, and Fischer (1972) - California
Drake, Cacchione, and Karl (1985) - California
Inman (1953) - California (La Jolla)
Inman (1957) - California
Inman and Risnak (1956) - California (La Jolla)
Inman, Swift, and Duane (1973) - Washington
Kachel (1980) - Washington
Larsen (1982) - Washington
Osborne and Yeh (1991) - California
Osborne, Yeh, and Lu (1991) - California
Pilkey and Field (1972) - Southeast United States
Shepard (1950) - California
Shepard and Inman (1951) - California (La Jolla)
Smith and Hopkins (1972) - Washington, Oregon
Sternberg (1972) - Washington
Sternberg and McManus (1972) - Washington
Sternberg and Larsen (1976) - Washington
U.S. Department of Commerce (1984) - California; Nearshore Sediment Transport Study
Vernon (1965) - California

North American Atlantic

Bowen (1980) - Canada
Butman, Noble, and Folger (1977) - Mid-Atlantic Coast
Figueiredo, Sanders, and Swift (1982) - Central Atlantic Coast
Gadd, Lavelle, and Swift (1978) - New York
Green et al. (1988) - North Carolina (Duck)
Greenwood and Mittler (1984) - Canada
Hall and Herron (1950) - New Jersey
Hubbard (1992) - U.S. Virgin Islands
Kraus, Gingerich, and Rosati (1989) - North Carolina (Duck)
Ludwick (1977) - Virginia
McClennen (1973) - New Jersey
Pearson and Riggs (1981) - North Carolina

Pilkey (1968) - Southeast Atlantic United States
Pilkey and Field (1972) - Southeast Atlantic United States
Reineck and Enos (1968) - Florida
Richmond and Sallenger (1985) - North Carolina
Stauble (1992) - North Carolina (Duck)
Stauble, Garcia, and Kraus (1993) - Maryland
Stubblefield, Permenter, and Swift (1977) - New York
Swift et al. (1981) - New York, Maryland, Massachusetts (Nantucket)
Swift, Thorne, and Oertel (1986)
Twichell (1983) - Georges Bank
Wright et al. (1986)
Wright et al. (1991) - North Carolina (Duck)
Vincent, Swift, and Hillard (1981) - New York
Vincent, Young, and Swift (1982) - New York
Vincent, Young, and Swift (1983) - New York
Williams (1976) - New York
Williams and Meisburger (1987) - New York
Windo, and Gross (1989) - Southeast Atlantic Coast

United States Gulf of Mexico

Bernard, LeBlanc, and Major (1962) - Texas
Brooks (1983) - Texas
Dupre (1985) - Texas
Hayes (1967a) - Texas
Hayes (1967b) - Texas
Hayes (1967c) - Texas
Hill and Hunter (1976) - Texas
Keen, T.R., and Slingerland, R.L. (1993a) - Texas
Keen, T.R., and Slingerland, R.L. (1993b) - Texas
Morton (1981) - Texas, Louisiana
Morton (1988)
Snedden, Nummedal, Amos, (1988) - Texas

North American Great Lakes

Osborne, P.D., and Greenwood, B. 1992 - Lake Huron

North Sea

Aagaard (1988)
Arlman, Santema, and Svasek (1958)
Morton (1981)
Swift et al. (1981)
Winkelmolen and Veenstra (1980)

Other Locations

Beydoun (1976) - Eastern Mediterranean Sea
Boyd (1981) - Southeast Australia
Channon and Hamilton (1976) - Southwest England
Cowell et al. (1983) - Southeast Australia
Figueiredo, Sanders, and Swift (1982) - Brazil
Gagan, Chivas, and Herczag (1990) - Southeast Australia
Gao and Collins (1992) - China
Hino, Yamashita, and Yoneyama (1981) - Japan
Jago and Borusseau (1981) - France
Kuo, Su, and Liu (1980) - Japan
Kuo et al. (1987) - Japan
Pae and Iwagaki (1985) - Japan
Roy and Stephens (1980) - Southeast Australia

Sedimentation/Stratigraphy References

Numerous studies are concerned with stratigraphy and sedimentology of the nearshore shelf. A lot of these studies are from coastlines with different wave, tide, and morphologic settings. References concerning the sedimentation/stratigraphic characteristics of onshore-offshore sediment transport are broken down by region (The sedimentation/stratigraphy reference list of Appendix B contains additional references related to this subject.):

North American Pacific

Bernard, Le Blanc, and Major (1962)
Cacchione et al. (1984) - California
Clifton (1976) - Washington, Oregon
Clifton, Hunter, and Phillips (1971) - Oregon
Dingler (1974) - California
Dingler and Inman (1977) - California
Greenwood and Mittler (1984) - Canada
Harms, Southard, and Walker (1982) - California
Harms, Southard, and Walker (1982) - Oregon
Howard and Reineck (1981) - Canada
Hunter, Clifton, and Phillips (1979) - Oregon
Inman (1957) - California
Komar, Neudeck, and Kulm (1972) - Oregon
Komar and Miller (1975)
Miller and Komar (1980)
Nittrouer and Sternberg (1981) - Washington
Pilkey et al. (1972) - Oregon

North American Atlantic

Brown, Ehrlich, and Colquhoun (1980) - Southeast Atlantic Coast
Davidson-Arnott and Greenwood (1974) - New Brunswick, California
Davidson-Arnott and Greenwood (1976) - New Brunswick, California
Duane et al. (1972)
Eames (1983) - North Carolina
Figueiredo et al. (1981)
Figueiredo, Sanders, and Swift (1982)
Green et al. (1988) - North Carolina (Duck)
Harms, Southard, and Walker (1982) - Georgia
Howard and Reineck (1972)
Gorsline (1963) - Eastern United States
Greenwood and Hale (1980) - New Brunswick, California
Greenwood and Osborne (1991) - New Brunswick, California
Howard and Reineck (1972) - Georgia
Howard and Reineck (1981) - Georgia
Luternauer and Pilkey (1967) - North Carolina
McBride and Moslow (1991)
Mearns, Hine, and Riggs (1988) - North Carolina
Meisburger and Judge (1989) - North Carolina (Duck)
Meisburger and Williams (1987) - North Carolina
Riggs (1979) - North Carolina
Riggs and O'Connor (1974) - North Carolina
Schmittle (1982)
Schwartz, Hobson, and Musialowski (1981) - North Carolina (Topsail Beach)
Shipp (1984) - New York
Snyder, Hoffman and Riggs (in press) - North Carolina
Snyder et al. (1993) - North Carolina
Stubblefeld, Paramenter, and Swift (1977)
Swift and Freeland (1978) - Mid-Atlantic Coast
Swift, Freeland, and Young (1979) - Mid-Atlantic Coast
Swift, Thorne, and Oertel (1986) - Mid-Atlantic Coast
Uchupi (1963) - Eastern United States
Uchupi (1968) - Eastern United States
Uchupi (1970) - Eastern United States
Wright et al. (1991) - North Carolina (Duck)
Wright (1993) - North Carolina (Duck)

United States Gulf of Mexico

Gorsline (1963) - Southern United States
Hill and Hunter (1976)
Morton and Winker (1979) - Texas
Nummedal and Snedden (1987) - Texas

North Sea

Aagaard (1988)
Reineck and Singh (1971)
Harms, Southard, and Walker (1982)

Other Locations

Clifton (1976) - Southeast Spain
Engstrom (1974) - Lake Superior
Field et al. (1981) - Bering Sea
Field and Roy (1984) - SE Australia
Figueiredo, Sanders, and Swift (1982) - Brazil
Flemming (1980) - South Africa
Greenwood and Osborne (1991) - Georgian Bay
Harms, Southard, and Walker (1982) - South Africa
Hunter, Thor, and Swisher (1982) - Bering Sea
Short (1984) - Australia

REPORT DOCUMENTATION PAGE

Form Approved
OMB No. 0704-0188

1. AGENCY USE ONLY *(Leave blank)*

2. REPORT DATE
February 1995

3. REPORT TYPE AND DATES COVERED
Final report

4. TITLE AND SUBTITLE
Literature Review on the Geologic Aspects of Inner Shelf Cross-Shore Sediment Transport

5. FUNDING NUMBERS

6. AUTHOR(S)
J. Bailey Smith

7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES)
U.S. Army Engineer Waterways Experiment Station
3909 Halls Ferry Road, Vicksburg, MS 39180-6199

8. PERFORMING ORGANIZATION REPORT NUMBER
Miscellaneous Paper CERC-95-3

9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)
U.S. Army Corps of Engineers
Washington, DC 20314-1000

10. SPONSORING/MONITORING AGENCY REPORT NUMBER

11. SUPPLEMENTARY NOTES
Available from National Technical Information Service, 5285 Port Royal Road, Springfield, VA 22161.

12a. DISTRIBUTION/AVAILABILITY STATEMENT
Approved for public release; distribution is unlimited.

12b. DISTRIBUTION CODE

13. ABSTRACT *(Maximum 200 words)*

This report reviews literature concerning the geological aspects of inner continental shelf physical processes, sediment transport, and stratigraphy. Although surf zone and nearshore processes and sediment transport have been extensively addressed in the literature, inner shelf processes and sediment transport, particularly in the cross-shelf direction, are not well understood. Inner continental shelf processes and related cross-shore sediment transport between the beach and the inner shelf have important implications for engineering works such as beachfill design and dredged material placement.

Specific topics considered include: depth of closure and extent of sediment transport landward and seaward of this zone; processes that cause cross-shore movement of sediment; amount and physical characteristics of beach material lost to the offshore; long-term fate of sediment that has moved offshore; relationship between depositional structures and flow processes; the impact of episodic storms on sedimentation; and the importance of the geologic framework on the inner shelf. Discussions pertain to the relationships between sediment transport on the inner shelf and the concepts of equilibrium profile, depth of closure, and sedimentation and stratigraphic characteristics of the inner shelf.

14. SUBJECT TERMS

Cross-shore	Geologic framework
Depth of closure	Inner shelf
Equilibrium profile	Sediment transport

15. NUMBER OF PAGES
161

16. PRICE CODE

17. SECURITY CLASSIFICATION OF REPORT
UNCLASSIFIED

18. SECURITY CLASSIFICATION OF THIS PAGE
UNCLASSIFIED

19. SECURITY CLASSIFICATION OF ABSTRACT

20. LIMITATION OF ABSTRACT